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PRINCIPLES
OF
SEDIMENTATION

PRINCIPLES
OF
SEDIMENTATION

BY

W. H. TWENHOFEL

SECOND EDITION

McGRAW-HILL BOOK COMPANY, INC.
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1950

PRINCIPLES OF SEDIMENTATION

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PREFACE TO THE SECOND EDITION

Since the publication of "Principles of Sedimentation" in 1939, several important books bearing on the same subject have been published among which are "Modern Marine Sediments" edited by Parker Trask, "Sequence of Layered Rocks" by R. R. Shrock, and, as this revision goes to press, "Submarine Geology" by F. P. Shepard and "The Sedimentary Rocks" by F. J. Pettijohn. The subaerial sediments of the Mississippi Delta have been studied in detail by R. J. Russell and R. D. Russell and those of the Mississippi River flood plain by H. N. Fisk. Important contributions to our knowledge of marine sediments and sedimentary processes in the sea have been made by the explorations of the Woods Hole and Scripps Oceanographic institutions, and much work has been done in the study of marine sediments by students in Holland, France, and Germany. Many cores and samples of bottom marine sediments have been collected over several parts of the Atlantic Ocean, the Gulf of Mexico, off the coast of California, in the Gulf of California, and about the East Indies. Several coral islands in the Pacific Ocean have been drilled and have yielded cores of the sediments penetrated. Many workers, in articles published over the past decade, have added to the knowledge of sediments and sedimentary processes. This revision attempts to present the progress that has been made. The number of illustrations has been nearly doubled.

In preparation of this revision the author has had access not only to the library of the University of Wisconsin, but also to those of Yale University and the University of Tulsa. The opportunity to use their resources is gratefully acknowledged.

W. H. TWENHOFEL

ORLANDO, FLA.
March, 1950

PREFACE TO THE FIRST EDITION

Early students of geology gave much attention to sedimentary rocks, little to sediments and sedimentation. Most early explanations of origin were based on observations of rocks, few on observations of sediments, and environmental factors were only incidentally considered. There was a strong tendency toward generalization and toward reference of all varieties of each group of sedimentary rocks to a single mode of origin.

During the last quarter of the nineteenth century entrance into that field of geology now known as sedimentation was largely through the channel of stratigraphy. Many, if not most, prominent students of stratigraphy of the last half of that century began as amateur collectors of fossils; few had much knowledge of chemistry and petrology and all had less knowledge of the environmental conditions controlling the deposition of sediments and the distribution of the organisms enclosed in fossil form in the sedimentary rocks. The goals were well-preserved and rare fossils. Characteristics of the containing rocks were of secondary importance, and little thought was given by most to the matter of environment.

The view of simplicity of origin of the sedimentary rocks began to be questioned before the close of the nineteenth century, and there were some who had become impressed with, or had had glimpses of, the many difficult problems connected with origins of the sedimentary rocks. Sir John Murray and his associates on the "Challenger Expedition" had become appreciative of these problems as they applied to marine sediments, and the difficulties were clearly visioned by such men as Johannes Walther in Germany and G. K. Gilbert in the United States.

The first decade of the present century witnessed the slow rise of sedimentation as a division of geology. Several geologists stand ahead of their fellows in appreciation of the importance of environmental factors in sedimentary processes. The many-sided A. W. Grabau and Joseph Barrell were among the first of these, and they seem also to have been among the first to devote considerable parts of formal courses to consideration of sediments and sedimentary processes. Their writings exhibit keen appreciation of the interrelations of the different kinds of sediments and a rather clear glimpsing of the parts played by environmental factors in production, transportation, and deposition of sediments. T. Wayland Vaughan—paleontologist, stratigrapher, and oceanographer—is another of that decade who showed an appreciation of the vastness of the then nearly unexplored domain of sedimentation.

There has been rapid progress since the early twenties. Several students have assembled the data relating to the minerals of sediments. Many have contributed to some understanding of sedimentary processes, and environments have been questioned the world over. A voluminous literature concerned with methods of study has been published and many techniques of present common practice were not even known as recently as ten years ago. Sedimentation has an American medium of publication in the *Journal of Sedimentary Petrology;* in Germany there is *Geologie der Meere und Binnengewässer.* Attempts were made in 1925 and again more thoroughly in 1932 by the present author and collaborators to organize the knowledge relating to sediments and sedimentary processes in the first and second editions of the "Treatise on Sedimentation."

The arrangement and presentation in the present work depart from any previously used. Emphasis is placed on the concept that sediments are products of heritage and environments, the latter usually the more important. Logical approach to an understanding of sediments and sedimentary processes is thought to be best attained by complete consideration of those environmental factors that have major influence in production, transportation, deposition, and subsequent modification of sediments. The subject is thus entered, and consideration of products is postponed until environmental influences have been explored.

Important references of general character and references published since 1932 are given at the ends of topics or chapters. For more complete references, as well as for the factual data on which the many principles presented in this work are based, the second edition of the "Treatise on Sedimentation" (1932) should be consulted.

The present work should be considered no more than a step on the road to a better understanding of sediments and sedimentary processes. There is so much that is unknown that one should be humble. Only the margins of the unexplored domain have been invaded. It is hoped that this work may stimulate some to extend the advance.

W. H. TWENHOFEL

MADISON, WIS.
September, 1939

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CHAPTER I

INTRODUCTION

The rocks composing the outer crust of the earth are either of igneous or sedimentary origin, or they are the metamorphic equivalents of these two kinds of rocks.

Igneous rocks form through cooling and solidification of rock in liquid form, this taking place either on, or beneath, the surface. In the latter case, the rocks may ultimately be brought to the surface through removal of overlying rocks by erosion. Igneous rocks at the surface are exposed to agents that seek to destroy them. Parts go into solution in the water that flows over them, or through cavities of various kinds that penetrate them; parts decompose and form new minerals of which some may be soluble and hence are removed in the waters responsible for decomposition; and other parts are broken from the parent rocks in one of the several ways in which it may be done. The products resulting from destruction are varied; but they may, in general, be placed in the three divisions of carbonates, clays, and sands, the last including gravels and larger fragments. These products are moved as they are separated from parent rocks and enter into long or short transportation. Short to long halts may many times interrupt the journey from places of separation from parent rocks to places of final deposition. In course of time the particles attain permanent deposition with more or less sorting or separation of the different constituents. The particles after deposition may become coherent through cementation, compaction, or recrystallization, thus forming rocks of the sedimentary group.

The new rocks may be raised by diastrophism to a level above that of deposition and placed under such conditions that they are again broken and destroyed; the materials are again transported, deposited, and again indurated to form other sedimentary rocks; and this sequence of events may be many times repeated.

The history may take another course. The sedimentary rocks may become deeply buried and subjected to great pressure, or great heat, or both combined, and in course of time be changed to form unlike the originals. These are the metamorphic equivalents. These may be uncovered by erosion, subjected to agents of destruction, and taken into solution as far as possible, and that which remains after solution may be transported physically to sites of deposition. These particles again may become indurated to form sedimentary rocks.

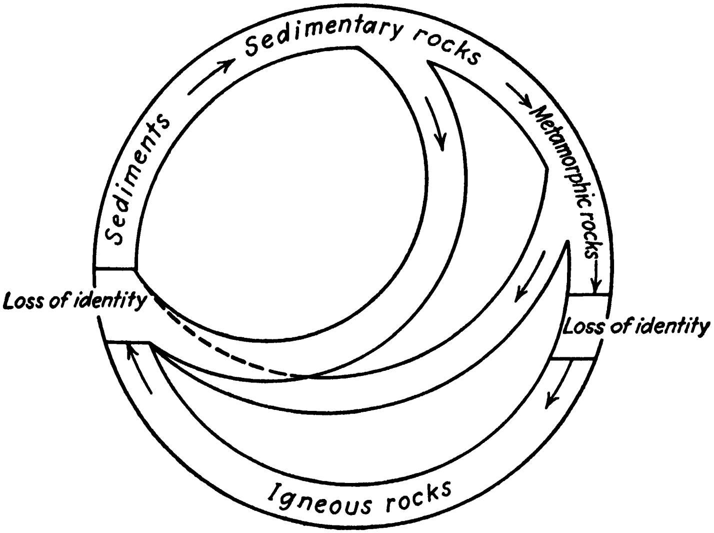

FIG. 1. The cycle of metamorphism, or rock transformation. The change is indicated as clockwise, and it will be seen that there are two places of loss of identity. The loss of identity indicated on the right is through reduction of the rocks to liquid form, the one on the left through destruction by disintegration and decomposition.

The metamorphic equivalents may not be brought to the surface, but may be subjected to such great heat as to pass into a liquid condition, and this liquid on cooling and solidification forms an igneous rock.

This sequence of events is termed the cycle of metamorphism. It is shown in diagram in Fig. 1.

The complete expression of this cycle has two places where identity of a rock is lost and a new rock is produced from the products of destruction. One place involves the destruction of an igneous or other rock as the initial stage in the preparation of a sedimentary rock, the other the destruction of the metamorphic equivalent of a sedimentary rock as the initial stage in the formation of an igneous rock. The cycle may be shortened so that there is but one loss of identity, that is, an igneous rock may be fused to return to the liquid condition. It may be greatly lengthened by the repeated destruction of a sedimentary rock, or its metamorphic equivalent, to form another sedimentary rock, and this sequence of events may be indefinitely repeated.

It is a matter of general knowledge that the exposed rocks of the earth represent progress in a part of the cycle, that igneous rocks have been repeatedly destroyed to produce the materials for the formation of sedimentary rocks, that sedimentary rocks have been destroyed to give the materials for new sedimentary rocks, and that metamorphic equivalents of both igneous and sedimentary rocks have been destroyed to yield materials for new sedimentary rocks. Every sedimentary rock is evidence of the partial completion of the complete cycle, and each sedimentary rock containing fragments from a preexisting sedimentary rock proves partial completion of the cycle. There are good reasons for believing that the entire cycle has been many times carried to completion.

That part of the cycle beginning with initiation of detachment of particles from rocks of any kind, through transportation and deposition, and finally induration or cementation through processes that approximate those of the surface, represents sedimentation. The other part of the cycle is concerned with the formation of igneous rocks and the metamorphic equivalents of both groups of rocks. Sedimentation is thus defined as representing that group of processes that is concerned with the destruction of any kind of rock, the transportation and deposition of the products of destruction together with materials derived from any source whatever, and the cementation or induration of these products so far as such is done by processes that are normal to the surface. Sediments are the materials with which sedimentation is concerned. They are the materials derived from any source whatever: rocks, organic matter, volcanic materials, and particles from outer space; transported in any way and in any medium from the places of origin on the earth, or where they appear on the earth, to the places of deposition, to be deposited under conditions of pressure and temperature that approximate those normal to the surface. Some transportation must take place, but it may consist of no more than detachment from parent rocks. It must, however, have been enough to have destroyed textures and structures of the parent rocks except as these may be preserved in the fragments. Rocks that decay without essential modification of texture, as is so frequently the case in the decayed gneisses and schists of the Appalachian Piedmont Region and elsewhere, are not considered sediments, as the particles are still in essentially original positions. Parent rocks may be either igneous or sedimentary, or the metamorphic equivalents of either of these two groups of rocks. Sediments may be derived from sources other than the rocks of the crust. Large and small particles reach the earth from outer space and may be deposited directly from the atmosphere in the midst of many kinds of sediments. Small particles may almost immediately enter into long transportation; large particles may remain near places of arrival for a long time. Waters derived from magmas doubtless from time to time enter the outer crust or even reach the surface to which they may contribute material in solution or colloidal suspension. Volcanoes at times eject great volumes of large and small particles. These may also become sediments. Plants and animals build or produce structures and materials of various kinds which may become sediments.

The common medium of transportation may be water, air, or ice. The manner of transportation may be by solution in the medium, by floating in colloidal or visible suspension, or by simple rolling or sliding. Deposition may take place on the surface or in the outer crust, and the causes may be physical, chemical, or organic, or some combination of these. The lower limits of deposition in the crust cannot be stated except that conditions of temperature and pressure should be such as to approximate those normal to the surface. This eliminates from consideration the deposits made by highly heated waters, although the causes of deposition and the reactions involved may not be greatly different from those taking place in cooler waters, except that there is little likelihood in highly heated waters of deposition due to organisms. Chemical reactions and supersaturation are causes of deposition in both places, and the products may in some cases be the same. There is no sharp boundary between the deposits made in waters of the two extremes of temperature. Introduction of the condition of temperature approximately normal to the surface eliminates the deposits made by precipitation connected with the cooling of lavas, although it is admitted that the causes of precipitation in the two mediums may be the same, and in some cases the minerals are the same.

Various processes are concerned in sedimentation. The detachment of particles from parent rocks includes processes of which some are chemical and others physical. The chemical processes are termed decomposition, rock rotting, or corrosion, terms meaning much the same thing. Both organic and inorganic agencies participate in producing the result. The most important chemical processes are oxidation, carbonation, and hydration. Others of less importance are chlorinization, sulphurization, and nitrification. The physical processes are generally termed disintegration or rock breaking. This is done in many ways, of which some are due to physical causes and others to chemical. Physical causes are found in expansion and contraction due to changes of temperature; the wedgework of ice caused by the freezing of water; the wedge effect produced by the development of new minerals connected with decomposition; breaking produced by impact, abrasion, and grinding; and, in some regions most important, breaking caused by movements of blocks of the earth's crust.

The agents of transportation are the currents of water, air, and ice; organisms; and simple gravity. Movement may be so slow at one extreme as not to be observable and at the other many miles may be traveled in a single day. Transported sediments may travel many hundreds of miles from places of origin, or the travel may be back and forth along a beach and end near places of beginning.

Deposition may be due to physical, chemical, or organic causes. Visible particles moved by water or air progressively come to rest as velocity of currents is decreased, and all visible particles except the very smallest settle on complete cessation of movement. The invisible particles in the air seem to have to be washed out. The very small particles transported in water settle only after they have united to form particles sufficiently large to settle under the influence of gravity. Particles that roll or slide cease motion when the resistance to rolling or sliding exceeds the pull of gravity.

Sedimentary rocks have certain features in common with igneous rocks. As liquid rock falls in temperature it may become saturated with some of the substances in solution. These are then precipitated. The same holds for waters with temperatures normal to the surface, and the processes are the same, the only difference being in the liquids. A volcano may eject vast quantites of material in the form of "ash" or larger particles. If these particles return to the surface sufficiently hot to cohere, the rock should be termed an igneous rock, but, if before reaching the surface they have cooled so as not to cohere, they are sediments.

Sedimentary rocks, as well as all others, are adaptations to environmental conditions. Each sedimentary particle formed on the earth, whatever its character, forms because environmental conditions permit it to form, and it persists or disappears as environmental conditions dictate. If an environment always remained constant at the conditions that permitted a mineral to form, there would be no changes. There are probably no places on the earth's surface, or in the outer crust, where immutability is possible. The aggregations of sediments are also due to environmental conditions and likewise the induration or cementation. Environments are constantly changing and with these changes in the environment, but with a lag, go changes in the sediments and the rocks they form. The rocks from which the sediments are derived disintegrate or decompose because of environmental conditions. These conditions determine the character of the particles produced by disintegration and decomposition. These particles begin transportation; this also is a product of environment. As transportation progresses from one environment to another, the particles attempt to keep pace, and each environment produces some impress. Each successive impress may efface or mar the effects of the previous impress, and the first impress, that of the environment of production, may be entirely effaced by long transportation. The impress of the environment of deposition may entirely efface those of the environments of origin and transportation, or more or less conceal these impresses so that they become of the nature of palimpsests to be deciphered with great difficulty.

Sediments are derived from five sources, which in order of probable quantitative importance are: (1) Destruction of preexisting rock. Sediments from this source are termed terrigenous sediments. (2) Structural remains of animals and plants. These are termed organic sediments and consist of the tissues, shells, bones, and excrements of animals, and the logs, branches, stems, and leaves of plants. (3) Clastic materials ejected from volcanoes which after a short or a long time settle upon the earth or in water. These are termed volcanic sediments. (4) Lavas and magmas may yield to their surroundings large quantities of highly heated waters, which may carry much and varied material in solution or colloidal suspension. Materials of this origin have been termed magmatic sediments. (5) Particles of extraterrestrial origin now reach the surface of the earth in large quantities, and no doubt have done so since the beginning of geologic history. These are meteoric or cosmic sediments.

The products of sedimentation are as varied as the environments to which they owe their origin. For purposes of consideration they must be placed in some classification. This is not easy, as there are all gradations between every form of sediment and all others. No single criterion can be used as a basis for classification.

Names applied to sedimentary rocks have many origins. Some relate to composition, as limestone; others to sizes of particles, as sand, gravel, etc.; and others relate to structure, as shale. Many terms have been used with several meanings. A more systematic terminology has been proposed but has been little used.

The first characteristic used for classification in this work is based on the method of transportation, whether transported in a physical or a chemical way. Sediments transported in a physical way form the clays, silts, sands, gravels, and boulders. The sediments transported in a chemical way are directly or indirectly responsible for the limestones, flints, rock salts, etc.

The subject of sedimentation is considered in successive chapters as follows: the factor of environment; production of sediments; transportation and deposition of sediments; products resulting from the operation of sedimentary processes; and the structures, textures, and colors of sediments.

Conditions and processes affecting sediments and sedimentation were probably the same in the past as at present, but at any place and time in the past intensities and concurrences of conditions and processes were rarely, if ever, the same. Thus, the sedimentary products resulting from conditions and processes have varied at any place with time.

A number of terms will be used more or less extensively throughout the book. It seems best to define these at the outset.

Base level of erosion. The base level of erosion is the lowest level to which any agent or combination of agents can reduce a surface. The base level of erosion of running water is a level that coincides with sea level at the shore of the sea and rises gently therefrom. The base level of erosion of aeolian agencies is some level that may be above or below sea level. The level is largely determined by the position of the water table. The marine base level of erosion is the lowest level to which marine agencies can cut a bottom. This is different at different places and rises from outer waters gently to the shore to join the base level of erosion produced by running water. There are many local base levels for which local conditions are responsible.

Base level of deposition. The base level of deposition is the highest level to which a deposit can be built. There are many local base levels of deposition, owing to local conditions. The base level of deposition due to marine agencies coincides with the base level of erosion, so that if both were in process of development a single plane would result that would be the base level of deposition over places that are filled and the base level of erosion over places that are cut down.

Profile of equilibrium. A surface is said to be at the general level of the profile of equilibrium when erosion and deposition are so nearly the same that the surface is not being lowered by erosion or raised by deposition. In detail it is a surface so nicely adjusted to the waters flowing over it that when these are loaded to capacity they neither increase nor decrease their load. Since waters change both velocity and load from time to time, this definition also states that the detailed profile changes correspondingly, so that when velocity changes without increase of load a surface is eroded, and when velocity of a loaded current decreases, the surface receives deposits. The detailed profile is hence a transitory feature. In final development the profile corresponds with the base levels of erosion and deposition, but profiles of equilibrium may exist on bottoms far above these levels.

Distributive province. A distributive province refers to the environment embracing all rocks that contribute to the formation of a contemporaneous sedimentary deposit and includes the agents responsible for the distribution.

Provenance. Provenance refers to the terranes or parent rocks from which any association of sediments was derived.

Allogenic (allothogenic) minerals. Allogenic minerals are the minerals in sediments that were derived from preexisting rocks.

Authigenic minerals. Authigenic minerals are those developed in sediments.

Subaerial deposits. Subaerial deposits are made above and not beneath a permanent body of water.

Subaqueous deposits. Subaqueous deposits are made beneath the surface of a permanent body of water.

CHAPTER II

THE ENVIRONMENTAL FACTORS

GENERAL CONSIDERATIONS

Studies of sediments and sedimentary processes must not fail to consider the environmental factors. Each sedimentary particle, no matter how small or large, owes its existence (unless it comes from outer space or is ejected from a volcano), transportation, and deposition to terrestrial environmental factors that may have varied from extreme to extreme from the time of its origin to the time of its inclusion in a sedimentary deposit. This may be illustrated by considering (*a*) a particle detached by glacial ice from a Pre-Cambrian rock on the east side of the Glacier Park Mountains of Montana and transported down the Missouri River through the semiarid prairie regions of the Great Plains to the Mississippi River and thence through the rainy lowlands of Louisiana to final deposition on the delta or beneath the warm waters of the Gulf of Mexico; or (*b*) another particle detached from the same terrane on the west side of the same mountains and ultimately carried down the Columbia River to the Pacific Ocean; or (*c*) a particle detached in the deserts of central Asia from a terrane exposed on the north side of the Himalayas and transported down some river of Siberia to deposition beneath the cold waters of the Arctic Ocean; or (*d*) a particle detached from the same terrane on the south side of the Himalayas, transported down the Ganges, and deposited beneath the warm waters of the Bay of Bengal. The terranes and the regions of origin may be the same, but the subsequent histories and the places of deposition may be as unlike as possible, and the sediments on deposition may be altogether different. The profound influence of environment on production, transportation, and deposition of sediments necessitates examination of this influence in detail.

The factors that constitute an environment are many, and in most cases they are extremely difficult to evaluate. Both organic and inorganic factors usually are present. The less apparent factors, which in many instances may be passed over by casual observers as unimportant, may be those that decide the events that take place and the sequel, the products that are made, and the organisms that dwell in the environment. Environments may be so nearly alike that it is difficult to detect differences, but the differences in the products may be striking.

All organisms dwell under the conditions of the environment to which they are adapted, and they may be killed, or at least suffer terribly, if environmental conditions change, yet the factor that may be the deciding one may seem extremely unimportant. A lowering of the temperature of but a few degrees may eliminate reef corals, and a change in atmospheric humidity not appreciable to man may be destructive to many plants and animals. Most plants are extremely susceptible to environmental factors. Slight changes in the acidity or alkalinity of a soil may be the deciding ones in determining the presence or absence of many plants and the animals that feed on them.

An environment remains the same as long as the factors that compose it remain the same in character and intensity. Such constancy is unlikely for a very long time. If the intensity of any factor is changed, if any factor is removed, or a new one enters, the environmental conditions are no longer the same, and the change may result in the elimination or modification of one or more of the processes that existed in the original environment and certainly will introduce conditions that may lead to elimination of some organisms. Products will change in accordance with changes in the environment.

Each organism competes with others, and each animal is either a herbivore, a scavenger, or a predator. Predators, while seeking other animals for food, may be the prey of other predators, and scavengers eliminate organic matter if it is accessible. If a single element in this organic complex is eliminated, there is initiated movement that is reflected throughout, and ultimately a complete change in an organic assemblage may result.

Important environmental factors on the land are temperature, atmospheric pressure, light intensity, humidity, precipitation, character of the soil, nature of the surface, and the life. The organic and inorganic factors react on each other to determine the aggregate impress of the environment.

An association of environmental factors invites the entrance of certain forms of life. These appear and their presence changes the environmental complex. This may be illustrated as follows: A virgin forest has surface animals living on its bark and foliage; ground animals living on and in its roots; and fungi, mosses, and various parasitic and epiphytic plants on and in its wood, bark, leaves, and roots. The association is a complex one and generally is composed of many kinds of animals and plants. Man enters the scene, the forest is destroyed, the land is plowed, domestic animals and plants are introduced, and with these come the various organisms that depend on the domestic animals and plants for their existence. The original organic elements have been destroyed, a new association has been introduced, and only a single organism, man, was responsible for the change.

A field has been bared by cultivation and abandoned. While bare, it is exposed to the burning rays of the sun, the drenching and washing of rain, the impact of wind, and in temperate and colder regions to greater or less intense frost action. Certain plants appear and with them herbivorous animals and their commensals and parasites for which these plants are food or home. Predators follow the herbivorous forms. The environment has changed and the change permits entrance of other plants with their parasites and commensals, the animals that feed upon them and the predators of these. Again there is a new environment and the sequel is the same as before, and this continues on and on without end. The soil is finally covered with organic litter; it is protected from the sun, from drenching and washing of rain, from impact of wind, and from frost action. Concurrently with this development the ground organisms change, and each change makes mandatory other changes, and there is thus no ending to the sequence of environmental change. If the environmental factors are such as to produce sediments, or to deposit them, the characters of the sediments change concomitantly with the changes in the environmental factors.

This fact of constant change in the land environment should be fully appreciated, and there should be no minimization of the fact that whatever takes place inevitably compels some further change in the environmental setup and some change in the sedimentary products, all of which in turn compel other changes in the environment.

Important factors in the aqueous environment are temperature of the water, winds that blow over the water body, quality and quantity of the solids and gases in solution, turbidity, depth, light, pressure, circulation, character of the substratum, and the organisms, both those dwelling in the water and those dwelling in and on the bottom materials. Depth in part controls the quantity of light, which is also controlled by the turbidity, which in turn is dependent upon distance from land and bottom. Depth also controls pressure. The physical and chemical factors determine whether and which organisms may dwell in the environment, and once the organisms are present they may so modify the various factors as to produce conditions that may lead to their elimination and at the same time invite the entrance of others. Waters of any depth have a population adapted to that depth. But these organisms make deposits, and they may also be made through the entrance of sediments from without the environment. These deposits raise the bottom and change depth, pressure, circulation, light, and perhaps temperature. These changes may permit entrance of organisms that previously were just on the edge of adaptation to the original conditions and eliminate others that cannot endure the new depth and the conditions thereby determined. The entrance of the new organisms and the disappearance of some of the old may start a chain of entrance and replacement that may completely change the organic association.

Animals and plants become adapted to every environmental factor that is normal to the aqueous environment, and in some cases this applies to conditions that are below the temperature of freezing of fresh water. As long as the environments vary within narrow limits the animals and plants survive, but, if the swing passes beyond the normal, destruction of organisms may be catastrophic. Thus, in 1881–1882 a current of cold water was swept to areas off the coasts of New England and Nova Scotia over bottoms normally bathed by the warm waters of the Gulf Stream. The change was attended by wholesale destruction of life. The bottoms appeared as if devastated, and an important food fish known as the tile fish was brought almost to the verge of extinction. This destruction profoundly altered the biological factors of the region, which in turn must have had repercussions on the chemical and physical factors.

Before presenting a classification of environments and the characteristics of each it seems desirable to examine several important factors that have large bearing on sedimentary conditions and the formation of sediments. These are physiography, diastrophism, climate, fire, close pasturage, and cultivation. Fire, close pasturage, and cultivation excepted, these factors are somewhat interdependent. A change in physiography can hardly fail to influence climate. Diastrophism in raising and lowering land surfaces affects climate and physiography and places sea level in a different relation to the bottom. Climate controls precipitation in character and quantity, the nature and intensity of the winds, and the kinds and abundance of vegetation mantling the surface; and the combined influence of climate and vegetation has bearing on the character and abundance of sediments produced and placed in transportation and determines some of the characters present after deposition. Relief conditions stream and wind velocity and thus has bearing on the quantity and quality of sediments transported. The distribution of responsibility to these and other factors for the characters found in any sedimentary deposit is extremely difficult, but this distribution must ultimately be made if geologic history is ever to be fully understood. It is almost impossible to consider any of these factors without paying attention to others, and thus others must be introduced as each is considered. This is attempted on the pages that follow.

THE FACTOR OF PHYSIOGRAPHY

Physiography is a factor of major importance in sedimentation, largely because of its great influence in controlling stream velocity and the position of the water table, and, under certain conditions, as an important factor in determining climate. Under conditions of much relief, stream velocities are strong and waters have maximum capacities and competencies. They may seize the particles released from parent rocks long before there is any opportunity for mature decomposition, and the results, if transportation is short, are deposits of more or less coarse clastic sediments. The water table under conditions of much relief may be far beneath the surface under the high places, and there is thus a thick zone given over to vadose water and to prolonged leaching and removal of colloids and solubles. Under conditions of low relief the surface waters move sluggishly and have low competencies and capacities, the water table is close to the surface, and there is only a very thin zone given over to vadose water and weathering. The loads of water under these conditions consist largely of dissolved and colloidal matter, both may be small, and these may be all that are brought to sites of deposition. Clastic deposits are then rarely possible unless the materials are derived by wave erosion from shores of lakes and the sea.

The factor of physiography is considered from the point of view of the cycle of erosion, and three stages are given consideration (Barrell).

HIGH REGIONS OF PHYSIOGRAPHIC YOUTH

A high region in physiographic youth may be a plateau or a youthful mountain range. A plateau has its surface more or less furrowed by canyons of greater or less depth and width. The climate may be wet or dry. Streams tend to be fast flowing. Sediments derived from plateaus that support an indifferent growth of vegetation tend to have immature decomposition, and under most conditions waters and winds have little difficulty in obtaining loads. There is little likelihood of permanent deposition in the bottoms of the canyons, and the aqueous sediments ordinarily are transported to, or beyond, the borders and localized on adjacent lowlands, or over bottoms of adjacent bodies of water. This is exemplified by the sediments derived from the Colorado Plateau by the Colorado River and carried to its delta across the head of the older Gulf of California. Sediments are more or less immaturely decayed unless they were derived from terranes that had undergone mature decay in an earlier history. Dry plateaus yield considerable volumes of dust, which is carried leeward and widely spread over lands with more humid climates, where ordinarily the dust loses identity as it becomes added to other sediments or becomes a part of the soil.

If a plateau is covered with vegetation, as once was the Cumberland Plateau of the Appalachian Region, the story is somewhat different. Winds acquire no load, and the plant cover also makes it difficult for water to obtain clastic sediments. Deepening of channels of streams yields a considerable volume of sediments and undermining adds to the quantity. Few of these sediments are deposited within the plateau.

A youthful mountain range has a windward and a leeward side. The higher parts of the range may be so cold as to preclude much vegetation. Where temperatures permit, the windward side may be expected to receive sufficient moisture to support a vegetable cover. The leeward side has a less abundant growth, and it may be scanty. All parts of the upland above the level of effective plant protection are sites of more or less intense block and granular disintegration, and any fine material produced remains but a short time after detachment from parent rocks. Fresh surfaces are constantly subjected to frost action, insolation, and incipient decomposition and to some wear through impact and abrasion. These effects may extend downward on the leeward side from the summit to the base and may be more intensive on lower levels than above. However, there may be a level on the leeward side where the combination of temperature and precipitation may support a vegetable cover, whereas at a lower level the rise in temperature and lessened precipitation may be such as to permit little plant growth. The materials on the leeward side tend to be immaturely decomposed and are more or less rapidly transported by ephemeral streams. These streams pass into dryer conditions on their descent and tend to become smaller because of evaporation and soaking into the substratum of previous deposits, and they may ultimately disappear. The suspended and tractional loads are dropped near the base of the upland, there to build alluvial fans composed of immaturely decomposed materials of which the particles range from very small to very large. If a belt of vegetation exists between the base and the summit, this may serve to lodge considerable quantities of sediments and thus lessen the quantities carried to the base. However, any sediments that reach the larger stream channels may be expected to continue to the bottom.

Results tend to be different on the windward side. If the precipitation falls as rain not in the form of cloudbursts, vegetation would thrive where conditions permit, and thus the windward side might support a sufficient mantle to permit the rocks to decompose maturely, to hold the water as it falls, and to yield it gradually. Little water would then be given directly to streams. Streams, fed largely by springs and seepage, would carry little other than dissolved and colloidal matter from the protected areas. If no part of the upland was above the snow line, vegetation would extend to the summit, and streams would obtain little in the way of coarse sediments other than what might be torn from banks and bed and might slide in from sides. Under these conditions the waters from an upland might bring little to the foot of the upland that would remain there and might carry to a farther site of deposition the small suspended load and certainly the loads of dissolved and colloidal matter.

If the upland extends above the snow line, glaciers may develop in the heads of the valleys, larger on the windward than on the leeward side, and these would deliver to swollen streams of summer, loads of clastic material ranging in dimension of particles from rock flour to large boulders. These would be carried to the foot of the upland to build alluvial fans and cover floodplains.

Contrasting the deposits of the two sides gives results as follows: The gravels on the windward side tend to be better sorted than those on the leeward and tend to contain more resistant materials, owing to the fact that decomposition progresses farther on the windward side than on the leeward. Deposits over sites of deposition fed by streams draining the windward side contain clays, silts, and sands. Those of the leeward side form poorly assorted piedmont deposits.

Colors of the sediments vary with the extent of decomposition, the colors of the minerals of the original rocks, and the quantity of organic matter buried with the sediments. If an abundance of organic matter is buried with the sediments, they tend to be dark. This is possible on the windward side; it is not likely on the leeward side. If there is a scarcity of organic matter deposited with the sediments, they would have either the colors of decomposition, red under warm conditions, yellow under temperate; or the colors of disintegration, that is, the colors of the composing minerals. The former is likely on the windward side; the latter is more likely on the leeward side.

The postulate was made that glaciers covered higher parts of the upland. If conditions should arise to eliminate these, there would be extension of vegetation to areas released from ice. The disappearance of ice would result in decreasing the quantity of sediments contributed to streams. These would not be loaded to capacity and thus would reach the foot of the upland in a condition to acquire a load from the loose deposits previously made when glaciers were present. The materials would then be transported downstream to other sites of deposition—the floodplains and deltas. This sequence of events has taken place in the mountains of Montana and Wyoming and is splendidly shown on the east side of the Big Horn Mountains. During the Pleistocene the mountains had far more extensive glaciers than at present. These supplied the streams originating in them with vast quantities of silt, sand, gravels, and boulders. The streams reached the foot of the mountains with loads they were unable to carry over the lowlands because of decrease of velocity. There was thus built a great thickness of deposits flanking the mountains. The glaciers have largely disappeared; areas formerly covered by them have now become covered by vegetation. No longer is the large load provided in the uplands; streams reach the lowlands as clear water with unused capacity and competency, and as they flow over the former deposits the previously unutilized energy is applied to remove these. The once extensive deposits are now represented by remnants between which the streams flow essentially as clear water. If the mountains should again become covered with glaciers and the conditions of the Pleistocene were restored, deposition would again begin over the areas about the foot of the mountains and the new deposits would hold disconformable relationships with the old.

A somewhat parallel result may be caused by fire or any other condition that tends to remove a vegetable protection. A fire may burn off the vegetation on the windward side of an upland, thus baring residual deposits to attack, and these may then be swept away in a very short period of time to form a thick deposit at the foot of the upland, the finer and more maturely decomposed going first, and lastly the more or less immaturely decayed rock particles at the base of residual accumulation. The deposit would have a certain degree of fineness at the base and would be coarser at the top. The charcoal produced at the burning would be carried onward owing to its low specific gravity, and thus there would be little or no evidence of fire. This sequence of events may also take place on the leeward side, but it is not likely that the effects at the foot of the mountains would be so pronounced, because of the probable lesser thickness of residual deposits and also because of the probability that they would not have decomposed to such fineness as on the windward side.

Since the advent of man this feature has been more pronounced than before because of the probable greater prevalence of fires and also because of close pasturage and clean cultivation, each of which has somewhat similar effects. The deposits resulting from these causes may be seen over extensive areas about the Appalachians where once fertile valleys have been covered by coarse sediments that represent the last things removed from the now bare mountain slopes.

An increase or decrease in rainfall on either the windward or leeward side of an upland may be reflected in the sediments deposited at the foot of the upland. The slope may be at the critical point where precipitation and vegetation are in such balance as to maintain the maximum thickness of residual materials. If rainfall should increase, it might exceed the absorptive capacity of the vegetable and residual materials and thus be compelled to flow away on the surface, each time taking a burden and each time leaving the substratum with less capacity to retain the next fall of water. Ultimately the surface accumulations would be entirely swept away and deposited at the foot of the upland to make a deposit not unlike that made by fire, but with the difference that considerable organic matter would be likely to be buried with the inorganic sediments, which, if in large enough quantities, would either bleach them through reduction of the iron or color them with the parts not decomposed. Again the vegetation might be in such relation to small precipitation that it could barely maintain itself. Any decrease in rainfall lessens the efficiency of the vegetable cover with consequent increase in the quantity of sediments removed and the dimensions of the particles of the load.

Changes of temperature also have effects. A rise of temperature has an effect like a decrease in rainfall, as it increases evaporation and transpiration, whereas a fall of temperature has the effect of an increase in rainfall.

PHYSIOGRAPHIC MATURITY

Regions of physiographic maturity have very largely been reduced to slopes, and elevation has little climatic influence.

The sediments produced and deposited are essentially the same as those from low regions of physiographic youth. Regions of these characters may support a profuse, scanty, or indifferent plant cover.

The water table would be closer to the surface beneath the hills than in the youthful stage of the cycle, and correspondingly, the zone of vadose water and depth of decomposition would be smaller. As movement of underground water would be relatively slow in a comparatively thick vadose zone, opportunity for acquirement of a considerable load of dissolved and colloidal matter would be favored.

If a region has sufficient rainfall to support an adequate cover of plants, there is likely to be a continuous flow of underground water into the streams from springs and seepage. The streams ordinarily carry little suspended matter except following those times when precipitation has been beyond the absorptive capacity of soil and vegetable matter. The sites of deposition ordinarily receive deposits of suspended and tractional matter in small quantities only, and there may be deposition of relatively pure colloidal and dissolved matter. Suspended matter is not often likely to have dimensions greater than sand size, although at any time there may be slides into the streams that yield larger particles. Colors tend to be gray owing to reducing effects of organic matter buried with the sediments. If conditions of deposition are such that organic matter is not decomposed, it remains with the sediments and they have darkness of color proportionate to the quantity. Soils tend to remain at places of production; draining waters are rarely loaded to capacity; and the suspended loads acquired are likely to be carried to sites of relatively permanent deposition.

If a mature region is one with seasonal precipitation and the dry season coincides with the growing season, there is necessarily a reduction of vegetation. This results in exposure of residual materials and rocks, much removal of loosened substances from exposed places, frequent high waters in streams, loads that are often near the limit of capacity, and the bringing of large quantities of sediments to sites of deposition. Burdens of streams are frequently so great that some sediments may not be taken directly to sites of permanent deposition, but tarry for longer or shorter periods over floodplains of aggradation. The limitation of vegetation reduces the possibility of much reduction of the oxides and hydroxides of iron, and in temperate and tropical regions these may be deposited unchanged; and, if the organic matter buried with the inorganic sediments is small, they may remain in that form through lithification. Sediments deposited under these conditions tend to have colors ranging from gray to red. Black is not common.

The vegetation of an arid to semiarid region of mature physiography is generally scanty, and precipitation may fall in torrential fashion. Streams become flooded after each heavy rain and carry loads that may approximate capacity. Floodplains are apt to be aggraded, and large volumes of suspended and tractional sediments are carried to sites of both temporary and permanent deposition. The load of dissolved materials is also high. Arid regions frequently have drifting sands. The surfaces of the sand particles may become pitted by impact, and the particles may be rounded to small dimensions. When contributed to flooded streams these may retain the characters given them in drifting and after deposition retain them as evidence of their aeolian production.

Physiographic Old Age

In the old-age stage of the physiographic cycle the surface is subdued, and the water table in humid and subhumid regions is at, or very near, the surface. The vadose water zone and the zone of weathering or decomposition are correspondingly thin. Soils are rather thoroughly leached, are essentially wanting in plant foods, have acid reaction, and probably support a highly specialized vegetation. Water draining from such lands may carry little dissolved or colloidal matter because of nearness of the water table to the surface and the thinness of the vadose zone. Colloidal materials may consist almost entirely of silica and iron hydroxide. Carbonates of calcium and magnesium would be in small quantity because of early removal.

Streams flowing from physiographically old regions with humid climates move slowly in meandering channels and are rarely turbid from other than colloidal matter. The suspended and tractional loads are usually negligible but may be important in times of flood and occasionally from collapse of banks due to undermining. Suspended sediments consist largely of organic matter and products of mature decay. The deposits at times may consist of clays and silts, both containing more or less organic matter by which the sediments may be bleached, if deposition is slow, or darkened, if deposition proceeds so rapidly that the vegetation cannot decay. The colloidal and dissolved materials give rise to flints (cherts), calcareous deposits, phosphates, etc.

If a physiographically old region is dry, there is little vegetation, erosion by water is greatly reduced, and winds are mainly responsible for removal of the products of decay. As winds remove only fine materials, there is concentration of the coarse products of rock destruction, and ultimately the surface becomes more or less covered with a veneer of lag gravels.

SUMMARY

The physiographic cycle has been assumed to produce a sedimentary cycle, but there are altogether too many variables to chance correlating what is observed in the geologic column with stages in the physiographic cycle. Distances of sources of sediments from sites of deposition and climatic conditions have much to do in determining the characters of the sediments that attain deposition. Sources of sediments long distances from sites of deposition may lead to deposits that give the impression of physiographic old age. A few hundred feet of elevation near sites of deposition may contribute far more coarse sediments than elevations of thousands of feet 1,000 miles away. Distance from sites of deposition need not be great to eliminate the coarse sediments, as illustrated by the Fraser River of British Columbia (Johnston, 1921), which leaves the mountains about 50 miles from the sites of deposition over its delta, takes its rise in high mountains, is fed by glaciers, but brings only fine sediments to its delta, and from the sediments one might assume that they were derived from a region in late topographic maturity. It may require only a few miles of transportation to eliminate most coarse materials.

Climatic conditions influence the character of sediments produced. Lands with lush cover of vegetation hold the products of weathering until decomposition is mature and yield sediments having the aspect of physiographic old age, whereas old lands that are dry and have scant vegetation but heavy rains at times may yield sediments to sites of deposition that give the impression of physiographic youth or maturity.

In correlations of cycles of sedimentation with the physiographic cycle it has been assumed that in the youthful stage the sediments are gravels and sands; sands, silts, and clays in the mature stage; and chemical and colloidal precipitates in the old-age stage. If a land were thoroughly protected by vegetation, there might be chemical and colloidal precipitates to the almost total exclusion of other sediments in the shallow waters even during the youthful stage; and, if the climate changed to dryer as the mature stage of the cycle was reached, the sediments might become silts, sands, and even gravels; and thus the sedimentary cycle, if interpreted in the conventional manner, would indicate the physiographic cycle incorrectly. Moreover, it must not be forgotten that for any sands and gravels there are equivalent silts, clays, and chemical and colloidal sediments deposited at the same time in other places, and that each of these is as fully representative of the youthful stage as any of the others. Change in position of sea level may deposit fine-grained sediments of the mature stage of the cycle over similar deposits of the youthful stage, and thus the vertical section would be subject to interpretation as indicating no change.

THE FACTOR OF DIASTROPHISM

Diastrophism is a major factor in sedimentation. It is diastrophism that brings a region to the condition of physiographic youth, and it may maintain it in that condition. There is a beginning to the diastrophic movement, a maximum, and a decline. As a region rises, the sediments yielded to the sites of deposition may be expected progressively to increase in coarseness and quantity to the time of maximum elevation, following which a decrease in coarseness and quantity would be probable. This concept is based on the assumption that other factors do not modify the results. The sediments consequent to this diastrophism are largely clastics and are termed orogenic. Typical orogenic sediments are considered to exist on the north side of the Alps in deposits known as the Flysch and Molasse. The production and deposition of the Flysch are referred to elevation of the Alps from initiation of the movement to the time of maximum elevation. It has been proposed to apply the term to all deposits of the same origin and same time relationship to orogeny. The Molasse lies above Flysch and is considered a deposit made subsequent to the time of maximum elevation. It is also proposed to use this term for deposits of the same origin and time relationship to orogeny (Van der Gracht). As orogenic sediments of the *flysch type* were deposited as the elevation was rising, they tend to increase in coarseness from the base upward and to be more or less deformed. If the elevation is of the nature of an overriding thrust block, some flysch deposits may be underneath the rocks from which the sediments were derived. *Molasse* deposits tend to increase in fineness from the base upward. As they were formed after deformation was complete, they are little if at all deformed, and they cannot possibly underlie rocks from which the sediments were derived. Parts of the Flysch of the Alps are actually underneath the rocks from which the materials of the Flysch were derived.

Orogeny has generally been assumed to be expressed in the sediments as coarse clastics, but, as shown on earlier pages, coarse clastics are usually not carried far from distributive areas, and unless the area undergoing orogeny had close by an area in position to receive and hold sediments, it seems improbable that any great thickness of coarse sediments would anywhere be present to attest the orogeny. However, orogeny stimulates erosion and, somewhere, distant sites of deposition would receive quantities of fine clastic sediments somewhat commensurate with the extent of uplift. Volume rather than coarseness of the clastic sediments should be a measure of uplift. Thus, such great thicknesses of shale or sandstone as are present in the Devonian of northeastern North America and in the Ordovician and Silurian of north central Newfoundland record uplifts of sufficient magnitude to produce the volumes of sediments concerned. Where the sediments are deposited, whether near or far from distributive areas, and the thickness of the deposits depend on the positions of the areas of deposition with respect to the base level of deposition.

Diastrophism is also responsible for fall of sea level, and it may also cause rise of sea level, although in some instances this is due to deposition of sediments in the sea. Apparent rise of sea level may also be caused by compaction of sediments.

SHORE-LINE PHYSIOGRAPHY

The nature of a shore line and the position of sea level, whether stationary, rising, or falling, have great influence on marine life and the character of sediments deposited in shallow water.

A shore line has influence both from the point of view of its irregularities and the elevation of the adjacent land above sea level. The elevation of the land above sea level very largely determines the quantity and the quality of the sediments deposited in the adjacent waters, and the irregularities determine variations. Both quantity and variation of sediments and irregularities of the shore line influence the forms and quantities of life that dwell in the adjacent waters. The nature of most shore lines very largely depends on diastrophism, or the relative rise and fall of sea level, as one factor and age as the other, although some shore lines owe their origin and characters directly to deposition or erosion. Shores due to movement of sea level may be either submergent or emergent. Submergence of a shore line results from relative rise of sea level (either because of diastrophism or filling of the sea with sediments) and brings waters over a land surface that may be in the youthful, mature, or old-age stage of the physiographic cycle. Emergence of a shore line places sea level in contact with what previously had been bottom. As many sea bottoms have rather uniform surfaces, such shore lines are generally regular, and thus both organic and inorganic depositional relations are more or less uniform and the waters are shallow for long distances outward from the shore.

Submergence of a land in physiographic youth sends waters inland up narrow valleys, and the shore line formed has deep indentations separated by more or less broad salients. As slopes of youthful lands tend to be steep, deep waters exist close to shore. Submergence of a mature land brings about somewhat like results except that indentations of water into the land are broader, the salients of the land are narrower, and the slopes of the sea bottom average less than in youth. Waves under either condition may vigorously attack a coast and produce currents that drag materials within the range of competency to bayheads and to bottoms offshore. Unless bayheads receive strong flowing rivers they become filled with sediments. The large particles torn from coastal rocks remain near places of production, usually headlands, until reduced to dimensions within the range of competency of the removing currents. Large quantities of less coarse sediments are carried outward from and along the shore. There is great variation in the character of the sediments deposited, both outward and parallel to shores, and in the organisms that dwell in the waters upon these sediments, as the environmental conditions are of wide range owing to differences in depth, wave action, turbidity, variations in the bottom sediments, etc. Places protected against the entrance of much sediment are likely to support luxuriant faunas and floras, whereas places of rapid deposition and bottoms swept by strong waves and currents may be almost barren of life. In the youthful stage of coasts of submergence of youthful and mature lands, large quantities of sediments are supplied, as the waves then have maximum power to erode; steep and perhaps high cliffs rapidly develop, and large quantities of sediments are made available through undermining. Large quantities are placed in transportation, and as competencies and capacities of the outgoing currents vary from day to day and even from hour to hour, it follows that the materials carried outward and down the slopes vary to an equal degree in quantities transported and in dimensions of particles. As bottom currents transport materials by traction and overlying waters in suspension, any checking of velocity results in deposition of the coarest tractional load as well as a part of the coarsest suspended load. The result is poor sorting.

Erosion is active on headlands and deposition in bays, particularly about bayheads and across bay entrances. Deposition across bay entrances may impede circulation of the bottom waters of bays so affected, produce development of toxic waters and deposition of black muds filled with hydrogen sulphide. Reduction of irregularities of the coast line follows, and as this is done there is reduction in the irregularities of the sediments deposited parallel to the coast line and some development of uniformity in the floras and faunas on the bottom so far as the variations are referable to coastal irregularity and conditions caused thereby. At the same time reduction of the slope of the bottom in the vicinity of the shore develops, and there is progression toward the development of a profile of equilibrium. Progressively the power of attack of the waves on the shore decreases as it becomes increasingly screened by shallow water; less and less sediments are acquired; they are moved around more and more on the shore before transportation outward; the outgoing currents become progressively weaker; and thus the sediments transported outward tend to average smaller in quantity and dimension. The tendency is for the sediments derived from the shore to become finer, but approach of the bottom to the level of the profile of equilibrium brings bottom sediments within the range of wave attack and leads to removal of some fine materials, so that the upper parts of vertical sections may contain only coarse sediments. In time the old-age stage of the marine cycle on a coast of submergence is attained, and the coast line has a configuration determined by lithologic structure and wave vigor. The coast is now effectively screened by shallow water over a wave-cut and wave-built platform, and the waves reach the shore shorn of power. The position of the shore is now far inland from that of youth; the profile of equilibrium on the seaward side rests on sediments eroded from the shore and on the landward side on a rock-floored platform. Many of the bayhead deposits of the youthful stage of the erosion cycle have been removed to be deposited outward on earlier offshore deposits. The vertical section of the sediments records the change that has taken place *in the somewhat progressive increase in coarseness of the sediments with progress upward.* The organisms of the earlier stages of the marine cycle may show much variation on each level from place to place, but in the upper levels they are likely to approach uniformity. If another submergence should take place following completion of the cycle, or at any time in the cycle, conditions would approximate those of the beginning, and thus the sediments would record in cyclic arrangement the conditions to which they were due. If any part of the bottom has reached the level of the profile of equilibrium before the new submergence occurred, that bottom would be covered with the coarsest materials that the waters were not competent to carry away. *These would be a part of the underlying unit and would be the last sediments deposited before the beginning of a new cycle.* The interval between the time that this bottom reached the level of the profile of equilibrium and the beginning of the new deposition would not be recorded, and the later deposits would have unconformable relationships with the old. The new sediments might be coarser or finer than the old, and *fine materials of the new sediments might fill interstices of the old* (Twenhofel).

Submergence of an old land produces a regular and even shore line and one bordered by shallow waters. The coast is not likely to be cliffed or to develop cliffs shortly after origin. Suspended and tractional sediments would not be common. Sediments brought to sites of deposition would be chiefly colloidal and dissolved materials. The surface submerged may have had a slope less than that of the profile of equilibrium, and therefore some erosion of the new bottom would be compelled. The development follows the sequence outlined for a coast of emergence.

A coast line of emergence brings a sea bottom above sea level and the adjacent areas of the old sea bottom to a position nearer sea level. As sea bottoms of shallow water generally have nearly level surfaces, elevation produces a more or less regular coast line. Cliffs are not likely, as the shore is protected by shallow water, and the sediments deposited offshore have character determined by the quantity of sediments that may be eroded from parts of the bottom above the base level of erosion, by the character of the land both in materials and in surface—that is, the stage in the cycle of erosion—and by the climatic conditions. The unemerged near-shore sea bottom is likely to have a slope less than that required for a bottom at the level of the profile of equilibrium. This necessitates erosion. This is done as follows: The waves break at some distance from the shore. This occasions erosion at the places of breaking with deposition shoreward of some of the materials removed from the bottom. These deposits may ultimately attain sea level and in many cases may be built above sea level to form barrier beaches or islands. After the deposits rise above sea level the winds may aid to raise the barrier to considerable height and width. Each barrier beach has a lagoon behind it. The lagoon serves as a trap to catch sediments derived from margining lands and brought in by tides from the sea. These deposits tend to fill the lagoon. They have a wide range in character in different lagoons and in some cases in the same lagoon. At the mouths of streams they may be sands, silts and clays, sands in tidal channels where these cross the barrier beaches, carbonaceous deposits where the waters of the lagoon are shallow and arenaceous and argillaceous sediments are not introduced in too large quantities, and calcareous sediments where depth, salinity, and distance from sources of clastic sediments permit. In course of time the bottom on the seaward side of a barrier beach is deepened so as to permit the waves to attack the barrier on that side. The landward side at the same time may be extended landward by sediments brought across the barrier by waves and winds. Thus, the barrier progresses landward through erosion on the seaward side and filling on the landward side. Ultimately the barrier progresses entirely across the lagoon, and its seaward shore coincides with the shore developed immediately following emergence. The lagoonal deposits have meanwhile been eroded to the base level of erosion, and the lagoonal deposits and those of the barrier are preserved only to the extent that they existed below that level. If a bottom on which deposits of a lagoon and barrier are made was above the profile of equilibrium at the beginning of deposition, *it is unlikely that any of these deposits would escape destruction.* More or less complete preservation of the barrier and lagoonal deposits may be accomplished if submergence takes place before the barrier has been moved back to the original shore. If extensive lagoonal deposits are found in the geologic column, the materials of the barrier responsible for the lagoon should be sought for.

The sediments acquired as the marine cycle on a coast line of emergence is in progress are deposited in the deeper waters where the bottom passes to a level lower than that of the base level of deposition. Here deposition may have been essentially continuous from the time before emergence began, but probably at different rates, as depths of water and environmental conditions determined by depth would have been different after the fall of sea level occasioned by emergence. This fall of sea level would have resulted in faunal and sedimentational differences in the later deposits, which would be expressed in an increased coarseness in the sediments deposited after emergence, as the finer materials originally deposited in the deeper waters would become overlain by sediments derived from sediments originally deposited in shallower water. The organisms entombed in the sediments would express the differences in depth so far as they were entombed in the places where they lived.

If a sea bottom has an irregular surface and a steep slope outward from the shore, emergence would not produce a regular and even shore line, but an irregular one. The sequence of events outlined above under such conditions would not take place. The sequence would be somewhat similar to that of a coast of submergence of a youthful or mature land.

Shore lines resulting from deposition have their environmental conditions determined by the facts of deposition. These are considered in connection with delta deposits and with coral and other reefs.

The preceding discussion is based on the assumption that the emergence or submergence is an accomplished fact, and no consideration was given as to what might take place as emergence or submergence was in progress.

Under stationary conditions of sea level a shore in process of erosion moves inland and leaves in its wake a wave-cut surface[1] which may be at the depth below sea level that approximates the base level of erosion. A part of the sediments produced by erosion of the shore is carried outward and deposited as bottoms are reached that are below this level, the coarsest at the shallower depths and the progressively finer beyond. As the bottom continues to be built upward, the sediments become progressively coarser toward the top, which on completion of deposition at any place is at the level of the base level of deposition. It is not to be assumed that the curve of increase in size of the sediments is a straight line or a smooth curve. There may be layers of fine sediments interbedded at many levels. When deposition has attained the base level of deposition, it comes to a close for as long a time as the conditions do not change. This may be a few years, a few hundred years, or many thousands, and of this time no record is made at that place. Should conditions change —sea level rise—deposition may be renewed, and the outer part of the former base level of deposition becomes covered with sediments and its surface becomes an unconformity.

A shore line migrates inland more or less rapidly for a rise of sea level. As sea level before a rise is assumed to have been stationary, a wave-eroded surface would have been developed in places. This is now placed in deeper waters, the base level of deposition is raised above the bottom, and hence deposits may be made on this wave-eroded surface. These deposits tend to be coarsest landward and finest seaward, and as sea level rises the sheets of sediments progressively extend farther and farther in a landward direction; the sediments laid down on the wave-eroded surface, *in general, are the coarsest in the section, and they become progressively overlain by finer sediments as the waters become deeper, so that the curve of particle dimension shows an increasing fineness upward with minor fluctuations of the curve consequent to the variable competencies of the transportation agencies from day to day.* If the rise of sea level is spasmodic, the base level of deposition may be attained in as many times, with the result that the section appears cyclic, and it may have numerous disconformities.

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[1] If one may judge from existing wave-cut surfaces eroded on relatively resistant rocks, these would be eroded but little below low-tide level. The widest wave-cut terraces with which the writer is acquainted are on the south side of Anticosti Island, and these are bare or covered with very shallow water at low tide. The rocks are limestones.

Much has been said in textbooks of geology of basal conglomerates. What seems to have been meant are marginal conglomerates, as they are really marginal to other sediments and lie at different stratigraphic levels in a section as seen over considerable areas and thus transect time diagonally. This conglomerate is the initial stratigraphic unit of the overlying strata. If sea level is stationary or retreating, there may also be a conglomerate, but it is the final deposit of the strata that lie below.

The conception of a basal conglomerate does not seem to represent the general situation. If sea level should rise rapidly, it may merely wash away the fine materials of the soils and leave the coarse, but this assumes a rise of sea level so rapid that the coarse materials cannot migrate inland with the shore. How rapidly sea level may rise is not known, but it may be assumed that it is usually not so fast but that there is abundant time to develop a wave-eroded surface beyond which there would be deposited the coarsest material that the outgoing currents were competent to carry. The very coarsest materials are likely to migrate with the beach, and, except for extremely large boulders, do not remain on the wave-eroded surface. This becomes covered from the seaward margin, and the basal (really marginal) materials are likely to be the coarsest in the vertical section, but it does not follow that these are gravels.

If sea level falls, the shore migrates outward, the former wave-eroded surface is progressively deserted, and a veneer of littoral sediments may be deposited over it with retreat of the waters.

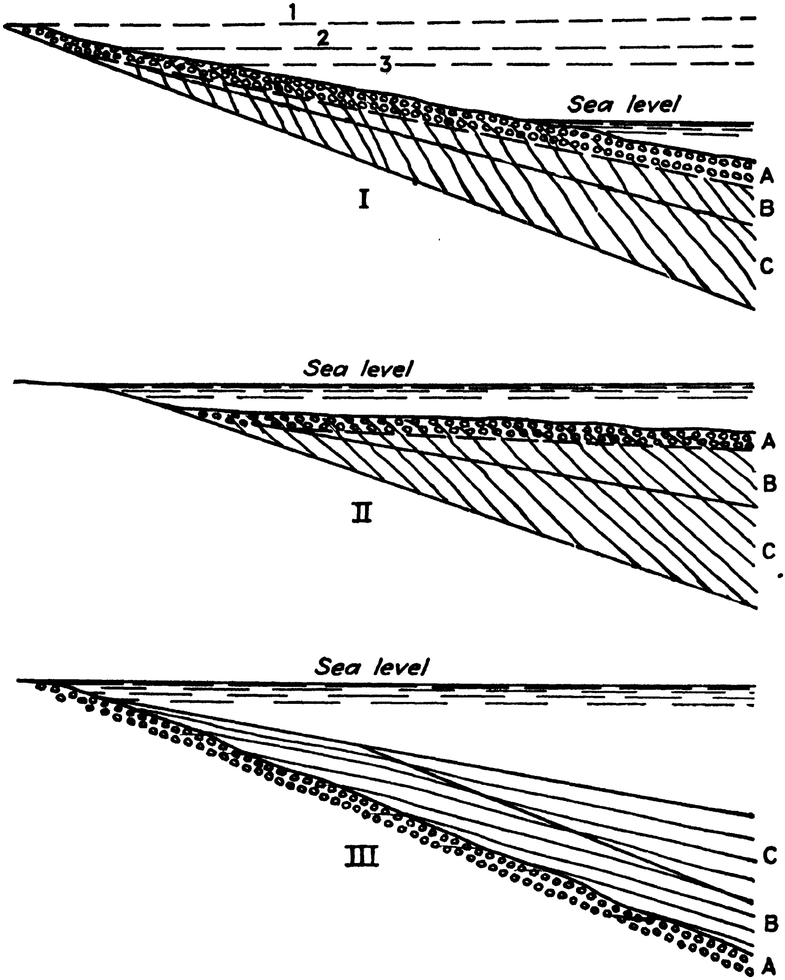

FIG. 2. Distribution of sediments over the shallow bottom of the sea under conditions of falling (I), stationary (II), and rising sea level (III). Depth to bottom is the only environmental factor considered.

I. 1, 2, 3 are former positions of sea level. *A*, *B*, and *C* represent lithic units. Bedding or time units cross the lithic units and are more steeply inclined than the lithic units. Coarsest sediments are at the top of the vertical section. Boundaries between lithic units are not sharp and they grade into those adjacent along the bedding units.

II. Figures and letters have the same meaning as in I. Lithic units are gently inclined seaward, but less steeply than the bedding or time units. Coarsest sediments are at the top of the vertical section if the bottom is at the base level of deposition.

III. Figures and letters have the same meaning as in I. Lithic units are inclined seaward at a greater angle than the bedding or time units. Coarsest sediments are at the base of the vertical section.

Deposits on the sea bottom above the base level of erosion are removed, and thus a part of the sedimentary record previously made is destroyed. As the top of the old deposits represents the coarsest of the section, these are carried outward and deposited above sediments that were made in deeper waters than exist under the new conditions. *The curve of particle size is thus likely to increase in average dimension upward with a sharp change and accelerated deposition at the level where sea level began to fall. No break is indicated by the increase in average coarseness. There may also be a change of fauna at this level, as the shallower waters may not be favorable to the organisms that have lived at each place when the waters were deeper* (Twenhofel).

The distribution of sediments with respect to size of particles, so far as distribution is controlled by depth, under conditions of falling, stationary, and rising sea level is shown in Fig. 2.

Under the conditions of a falling sea level the coarsest sediments, so far as distribution is controlled by depth, are at the top of the vertical section at any place [Fig. 2(I)]. The lithic units incline seaward at a smaller angle than the bedding or time units.

Under conditions of a stationary sea level the coarsest sediments are at the top of the vertical section. The lithic units are very gently inclined, the bedding or time units are more steeply inclined [Fig. 2 (II)].

Under conditions of a rising sea level the coarsest sediments are at the base of the vertical section at any place, the stratification or bedding units (time units) are less steeply inclined than the lithic units. This is shown in [Fig. 2 (III)].

Actual occurrences of some of the above relations have been described by Malkin and Echols in their two excellent papers dealing with conditions in the Tertiary of the Gulf Coast. Transections of time units by lithic units are excellently shown.

THE FACTOR OF CLIMATE

Climate seems to have its greatest impress at the places of production of sediments, but it aids in placing them in transportation and has influence on deposition, particularly with respect to continental sediments. The climatic impress in the production of sediments may be entirely obliterated in transportation, and the total climatic impress made upon the deposits may be entirely lost through diagenetic processes that modify the sediments after deposition. The effects are much more apparent in continental than in marine sediments, as waves, currents, organisms of the bottom, and the later diagenetic processes in the latter tend to eliminate any climatic impress, and marine sediments may carry little evidence of the climatic conditions concerned with production, transportation, and deposition. It of course follows that under any condition the organisms buried with the sediments reflect the climatic conditions under which the organisms lived. It must be known, however, that they lived where found, as dead organisms are often transported long distances from the places where they lived and become buried in sediments to which they may bear no climatic relation.

Certain broad generalizations are possible. The sediments produced by rock decay in tropical and near-tropical upland regions are dominantly red; those of the polar parts of the temperate regions are yellow to gray; and those of polar regions gray. However, the colors of the sediments of any region may be those of the parent rocks. Regions that are wet and have poor underground drainage tend to have dark soils above and light below. Prairies and steppes that are not so dry as to approach the condition of semiaridity tend to have black soils like the *tschernosem* of Russia and the *regur* of India.

The minerals that decompose or fail to decompose under continental conditions are intimately related to climate. Under conditions that do not permit complete or mature decay such minerals as feldspars, amphiboles, pyroxenes, and olivine may escape decay. Under more exacting conditions they would decompose to carbonates, clays, iron oxide, and quartz. In such regions as southern California, decay is rarely carried to completion, with the result that many minerals are found in the sediments which upon the southern Piedmont of the Atlantic Coast would have disappeared (Mackie, Reed).

The colors of sediments are closely related to climatic conditions, chiefly to the two climatic factors of temperature and rainfall. Under conditions of mature decomposition and solution the colors of the residual products of the many different kinds of rocks are essentially the same for each kind of climate, but under conditions of less complete decomposition the colors of the residual products may be greatly different and bear considerable relation to the parent rocks from which the sediments were derived. Hager states that sediments derived from the Cretaceous Taylor and Navarro formations of Texas are brown; whereas those from the Tertiary Wilcox, Mount Selman, and Cook Mountain formations have buff or red colors. As the climatic conditions are the same, it follows that the differences in color are due to differences in the characters of the original rocks.

Soils produced in one region have been carried to others in an endeavor to learn the effects of the climatic impact. The soil experiment stations of California, Kansas, and Maryland transported soil produced in each state to the others. The experiments covered seven years, a short time geologically, but the great differences resulting in even that short time are extremely noteworthy. The California soil was a distinct red brown. It became a darker red brown in Kansas and a dirty gray in Maryland. The Maryland soil was a light brown, but it became dark brown in Kansas and reddish brown in California. A black-brown soil of Kansas became light brown in California and light ashen gray in Maryland. These color changes were naturally connected with the chemistry of the soils, and analyses showed that changes in the content of silica, alumina, potash, lime, and iron oxide had taken place (Lipman and Wayrick).

Climate is responsible for vegetable growth, which in turn influences decomposition of underlying rocks. Vegetable materials fall to the ground to decompose there. Organisms carry a certain quantity into the ground, where decomposition may produce reduction of iron compounds, generally to carbonates. Waters charged with carbon dioxide from decaying organic matter remove carbonates, and the soils are leached of these. Vegetable growth restricts aqueous surface erosion and altogether prevents acquirement of any burden of inorganic materials by wind. With lush vegetation there is a large probability that considerable quantities are transported and buried with inorganic sediments. These reduce iron oxides in the sediments and produce the essentially colorless carbonate or silicate. Sediments originally red become blue to gray as a consequence.

The influence of climate in the production of soils and thus sediments is emphasized in the classification of soils by Glinka and Marbut, who consider climate as the most important factor in soil production and accord only minor importance to the character of the rocks from which the soils were derived. The ectodynamic soils—those due to factors external to the rocks—were placed in six classes corresponding to six types of climate. These are as follows:

1. Soils formed under optimum moisture conditions. These form under conditions of large quantities of moisture and a high mean annual temperature. Percolation is downward. These soils are low in humus, leached, and largely composed of oxides and hydroxides of silicon, iron, and aluminum. Colors range from yellow to red. The extreme expression of this type of soil is *laterite,* which is formed on well-drained areas under tropical and subtropical conditions.

2. Soils formed under average moisture conditions. Conditions are such that moisture and temperature permit accumulation of large quantities of unstable humus compounds. Percolation is downward. These soils are composed of the oxides and hydroxides of silicon and aluminum and are low in the oxides and hydroxides of iron. The soils are usually leached, and colors are light. They form under tropical and temperate conditions but are rare under the former. The extreme expression is *podsol.*

3. Soils formed under moderate moisture conditions. These soils develop in regions where temperature and moisture permit a lush grass growth, but which are not favorable for decomposition of organic materials. They are the soils of temperate latitudes. Percolation is downward. They contain much carbonate, sulphate, and humus. Colors are dark to black. The typical expression of this type of soil is the *tschernosem* of Russia.

4. Soils formed under insufficient moisture conditions. Percolation at times is upward. These have low humus and much carbonate and sulphate. Decomposition of the materials of the soil is rarely mature and colors range from gray to brown and red. These are the soils of the dry prairies.

5. Soils formed under excessive moisture conditions. The formation of these soils is favored by coolness of temperature. They have high organic content and are the black peaty soils of swampy conditions. Percolation is lateral.

6. Soils formed under temporarily excessive moisture conditions are the soils formed in regions of more or less semiarid climates. These are the so-called alkali soils and are high in such

readily soluble salts as sodium sulphate, sodium carbonate, and sodium chloride. They form under semiarid conditions subject to periods of seasonal rainfall. During the dry season the movement of underground water is upward.

It will be noticed that this classification places emphasis upon the presence of moisture. Consideration is given to temperature, but it seems to be relegated to second place. It seems probable that the two factors have a somewhat equal importance, since rainfall that might be optimum under one condition of temperature might be deficient under another. This is illustrated by comparison of the Gulf of St. Lawrence region with that of the region about Dallas and Fort Worth in Texas. A condition of average to excessive moisture exists in the former and the moisture is insufficient in the latter. Yet both have essentially the same precipitation. The factor that controls is temperature, which limits evaporation in the Gulf of St. Lawrence region and favors it in the Texas region. The diagram of Fig. 3 shows these relations, and it may be seen that for the same rainfall but different temperature the character of the soils may range from raw humus and podsols to sand, dust, and salt soils.

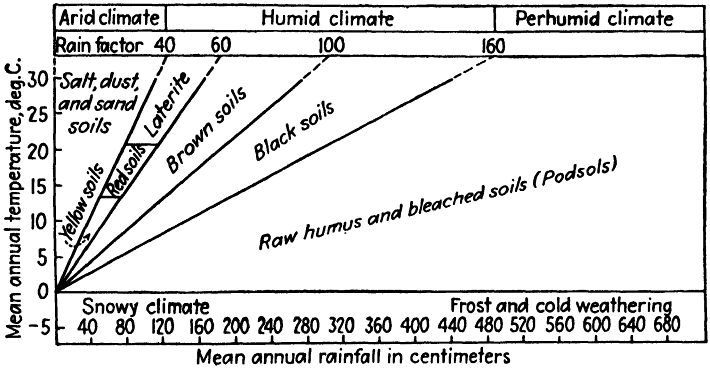

FIG. 3. Diagram showing relation between soils and temperature and rainfall. (*After R. Lang, Verwitterung und Bodenbildung als Einführung in die Bodenkunde, Stuttgart* (1920), 119.)

In the matter of transportation of sediments the factors of temperature and precipitation have somewhat parallel importance. High temperature with adequate precipitation supports a lush growth of vegetation and completely prevents atmospheric transportation of inorganic sediments and greatly limits the acquirement of a load by water. On the other hand, high temperature with insufficient moisture produces prairies, semiarid lands, and deserts with the attendant herds of animals, where it is not too arid, to eat the vegetation and grind the surface of the soil to dust. Acquirement of a load by both atmosphere and water is easy.

The form of precipitation has great importance in transportation, particularly in the acquirement of a load. If the precipitation is gradual, most, if not all, soaks into the ground, and there is little surface runoff and hence little removal of soil materials. If the precipitation is in the form of cloudbursts, absorption of only a small part of the water is possible, and large runoff with large removal of soil materials is a consequence. The times of precipitation are also important. If the water falls early in the growing season before the plant cover of annuals is established, the surface is more favorable for excessive runoff and removal of soil materials than at any other time.

For purposes of discussion, climates must be classified. The most convenient bases for classification, so far as sediments are concerned, are precipitation and temperature. On these bases, climates may be divided into frigid, arid, semiarid, climates with excessive moisture conditions, and climates of seasonal dryness. These five divisions are considered in the order named (Barrell).

FRIGID CLIMATE

The term frigid climate is considered to apply to regions that have the substratum a little below the surface permanently frozen and have a more or less permanent cover of snow and ice. The surface of the substratum may thaw during the summer season. A region of this kind may also be arid if the precipitation is so small that the surface is bare most of the time. Under the conditions of a frigid climate the destruction of rocks is almost entirely physical and is produced by insolation and the action of ice and frost and to some extent through abrasion, impact, and grinding. Plant growth has difficulty in establishing itself, and as a consequence much surface without snow is bare. The plants are lichens and mosses and various small forms of other phyla. The absence of plant protection gives opportunity to wind, and all fine material on the surface tends to be removed as rapidly as it becomes exposed. There is very little decomposition, and particles released from parent rocks are, to a large extent, fresh and unaltered. There is little or no oxidation of iron minerals and little to no coloring due to secondary iron oxides. Unstable minerals of all kinds may be released from the parent rocks and persist in the residual products. There is no underground water circulation, and the frozen rocks beneath the surface remain fresh and unchanged.

The colors of materials produced by breaking of rocks are those of the parent rocks except in cases where the rocks have been reduced to powdered form, in which case the color of the aggregate tends to be light.

The agent of transportation may occasionally be water, but wind and ice are those most prevalent. Water carries the materials to lower positions, perhaps the sea; winds carry them widely; and ice leaves them anywhere on the surface, takes them to the sea, or floats them away in icebergs. The waters of frigid climates are cold, and hence there is little chance for much decomposition of sediments either during transportation, at deposition, or after.

Materials transported by ice may be striated, and striations may be present on any surface over which ice moves.

ARID CLIMATES (DAVIS)

The term arid climates as used in this work applies to regions more or less permanently dry owing to deficiency of rainfall. The regions support little plant life. Rain is infrequent, and it may be in the form of heavy downfalls.

At first appearance it would seem that decomposition would not be prevalent in regions with arid climates, and most works on desert geology stress the view that the dominant processes of rock destruction are physical in character and that decomposition plays but a small part. This is evidently true only *if the rocks are not subject to decomposition.* If the rocks are decomposable, the traditional generalization does not seem to be true. Groundwater level of arid regions is far beneath the surface, and hence it is possible for air to circulate for long distances beneath the surface in the various openings of rocks. This air, before entering the ground, may be of high temperature with a low relative humidity but with absolute moisture content high. It becomes cool as it penetrates the rocks, and the relative humidity is almost certain to rise to the point that condensation of moisture takes place on the surfaces of cracks and other openings in the rocks. As both atmosphere and water are present, the result is decomposition with the consequence that *decomposable rocks become hydrated, carbonated, and oxidized for long distances beneath the surface.* Under ordinary conditions there is not enough water to remove the products of decomposition, and they remain at places of origin. Decomposition produces expansion of the materials affected, and this forces other parts of rocks apart. Thus, decomposable rock is reduced to fragments and the result is that decomposable rocks are masses of more or less rotten materials for great distances beneath the surface. The depth to which decomposition extends depends upon the depth of the water table and the facility by which air can penetrate the rocks. Such rocks as sandstone, limestone, and shale are little affected, as these rocks are end products of decomposition and shale is not readily penetrated by air; hence these rocks persist as prominent features of arid regions for long periods of time.

The general absence of water precludes much solution, and movement of the limited underground water may be upward and be evaporated at the surface. This concentrates soluble substances in, or on, the surface materials to form caliche. The result is that soils of arid regions are higher in soluble substances than those of moist regions and correspondingly lower in insoluble constituents.

The colors of the surface materials of arid regions vary with conditions. In many cases decomposition has advanced to oxidation of the iron, and the residual accumulations are brown or red. As a rule, decomposition is rarely so mature, and the dominant colors are shades of yellow or gray, corresponding either to the colors of the rocks of derivation, to the degree of decomposition, or to salts placed in the surface materials by waters evaporated at the surface.

As there is a scantiness of vegetation over the surface of arid regions, full play is given to the wind and it constitutes a most important agent of transportation. Small particles are swept away as rapidly as detached and exposed and in course of time are carried to other regions. Particles too large to lift are either left at the places of production or rolled elsewhere. Those remaining behind are designated *lag gravels*, whereas those rolled away are mostly sand and form dunes.

Water may also be an agent of transportation in arid regions, and sediments may be carried by water into desert areas from adjacent highlands. The occasional rains may be of such intensity that complete absorption of the falling water by mantling materials is not possible and each low area leading away from high areas becomes filled with torrents loaded with sediments. Temporary or permanent lakes may form in which deposits of evaporites may be made between times of deposition of suspended and tractional sediments.

Temporary lakes disappear with evaporation of the water, following which, rising underground water covers the surface with caliche or concentrates the salts in the surface materials as cement, concretions, or individual crystals. Under some conditions the crystallization of the salts in the sediments may completely disintegrate them and make them an easy prey to the frequent winds, and many depressions become deepened in the intervals of dryness. If the deposits of the temporary lakes are sufficiently tenacious, they become thoroughly mud-cracked in the intervals of exposure. Sands may drift to fill the cracks and lie beneath the raised edges of the polygons. Or the mudcrack polygons may collapse to form a breccia of mud fragments, some of which may be picked up and rolled on their edges to assume coin shapes of greater or less dimension.

Arid regions maintain small permanent animal populations. These should occasionally leave tracks in muds and sands.

SEMIARID CLIMATE

Regions of semiarid climate are defined as those that do not have an adequate protective cover of vegetation because of the combined impact of temperature and limited precipitation. The uplands usually support only a limited seasonal growth of annual, biennial, and perennial plants, which are of wide variety, but among which grass, yucca, members of the Leguminosae and Asclepiadaceae (milkweeds), and cacti are common. There is a prominent scarcity of arborial growth on the uplands, but low shrubs, as the prairie rose, small elms, and others, are not uncommon, and mesquite may thrive over valley bottoms of regions that are not too cold.

Many plants are spinose. Stream courses are usually defined by bands of trees, and so may also be the areas of exposure of certain geological formations. Rainfall tends to be seasonal and in large quantities over limited times. The plants flourish with the seasonal rainfall. They spring to blossom as if by magic to wither as rapidly with advance of the dry season. Haste seems to be the essence of the plant growth, as they must both blossom and mature their seeds in a relatively short time.

The ground-water table is relatively deep, and movement is apt to be upward during the dry seasons. There is likely to be much deposition of soluble and colloidal substances in and on surface materials. These are chiefly carbonates and sulphates of calcium, magnesium, and sodium, and they form *caliche*, *tepetate*, *kunkur*, *hardpan*, *alkali*, concretionary structures, and cement. The deposits may parallel the surface and may form a more or less case-hardened crust, the *duricrust* of Woolnough. This may attain a thickness of 5 ft. or more and be so firm as to be used for stone. Concretionary structures may be hollow, generally have irregular and peculiar shapes, and often have *breadcrust* surfaces. Greatest concentration of soluble and colloidal substances is over, or in, the materials of low and flat places. Under some conditions crystallization of the salts crumble the containing materials into particles small enough for wind removal. The frequent excess of surface waters permits considerable removal of the soluble salts, and streams flowing from semiarid regions generally have a large content of dissolved substances.

Decomposition may extend to depths equal to the lowest limit of the water table. Soluble rocks beneath the surface, as limestones, may become honeycombed by solution cavities, and rocks with soluble cement, as some sandstones, may have the cement removed so that they crumble to sands. Decomposable rocks, as granites, basalts, etc., become hydrated, carbonated, and oxidized to depths equally as great. The decomposition may not be complete, and it rarely reaches the stage of complete oxidation of iron compounds. Colors of soils tend to be light, but the range is from light to red with the common colors some shade of yellow to gray.

Regions of semiarid climates tend to have strong and frequent winds. These winds, owing to sparseness of upland vegetation over much of such regions, the presence of crumbled soils produced by crystallization of minerals brought by ascending waters, and the production and raising of dust by herds of animals, lead to much aeolian erosion, and many depressions without outlets are formed. These may be eroded at any elevation and become filled with water and populated by tadpoles during the rainy periods, at which times they also may be sources of water for land animals. The aquatic organisms die during the dry seasons and leave their skeletons on the surface of the mud-cracked deposits, on which may also be found the tracks and often the skeletons of land animals.

Rainfall tends to be concentrated, which, coupled with the limited plant growth, leads to excessive runoff. Streams respond flashily to rainfall, and channels that are without water one hour may become filled with raging torrents the next. Mudflows may be frequent owing to sparseness of vegetation.

Sediments from semiarid regions are not likely to contain much organic matter. They tend to be deposited over alluvial fans, floodplains, and deltas, where they form yellow to gray deposits that tend to be high in soluble matter. The deposited sediments become extensively mud-cracked during dry seasons, and the mud-cracked polygons may be so firmly cemented by contained colloidal substances as not to collapse on being wetted. They may also break up to form rubble, which on cementation becomes mud breccia. It is possible for salt and gypsum to have local deposition, but these are not likely to have much importance.

CONSTANTLY WET CLIMATE

Regions of constantly wet climate have a wet substratum throughout the growing season, and the substratum rarely if ever becomes completely dry. Constant rainfall or rainfall separated by short periods of time is necessary to produce this condition. A constantly wet substratum may be present in regions of other climates where a local high water table exists because of surface irregularities. Regions with constantly wet climates support a lush growth of plants, and the dead plant materials fall upon a wet substratum where in course of time they reach a state of semidecay and then function to lift the water level higher.

Surfaces of constantly wet regions may be flat or have considerable relief. The latter makes it more or less difficult to maintain wetness, as drainage is favored, and this makes for lowering of the ground-water table. Circulation of the underground water is rarely upward. Movement is mainly lateral, and there is no tendency for concentration of dissolved substances in the surface materials. The abundance of water together with an abundance of carbon dioxide and organic acids from decaying plant matter favors early solution and removal of soluble materials, so that those parts of the substratum through which there is circulation are rather thoroughly leached and thus are low in *potash, soda, lime, and magnesia.* There is also probable removal of all iron, as the decaying organic matter reduces the oxides to the carbonate, which may be removed in solution. Should sulphates be present in the soils, these may be reduced by bacterial action to hydrogen sulphide and ultimately to sulphuric acid. This would hasten removal of iron. There is also probable removal of much silica, as this is aided if waters are provided with organic matter. The result is that residual materials in constantly wet climates are composed largely of such insoluble substances as *hydrous aluminum silicates, quartz, and extremely resistant minerals inherited from parent rocks, and the more nearly the surface approximates base level the more nearly do residual materials approximate this composition.* Considerable depends, however, on temperature and relief. A constantly wet substratum is difficultly maintained in regions of high temperature and considerable relief; the same rainfall in a region of low temperature and the same relief may maintain a wet substratum. But low temperature retards decomposition and decreases solution, with the result that some soluble substances remain and decomposition may not be complete. Maintenance of effective protection through plant growth may be difficult on steep slopes, and fragments released from parent rocks may not remain sufficiently long to do more than incipiently decay. They may begin transportation shortly after detachment and ultimately be deposited little altered.

The slight depth of the ground-water table on level surfaces limits the depth to which decomposition may extend, with the result that the thickness and the quantity of the residual materials produced are not large. This is illustrated by the persistence of highly polished and striated surfaces made by Pleistocene glaciers in the northern hemisphere which have undergone little change in the some 25,000 years since the glaciers disappeared.

Colors of residual materials of regions that have constantly wet climates range from black to white. The materials at the top of the soil profile are mostly black, as they may consist in large or considerable part of semidecayed organic materials. The thickness of the black materials at the top varies with the conditions and is thickest where temperature, wetness, and relief favor plant growth and accumulation. The black materials pass downward into leached residual materials (*podsols*) which are some shade of gray. Sediments derived from these residual materials tend to be mixtures of organic and inorganic substances and to have some color in the range from black to gray. Effects of desiccation may, but should rarely be present in deposits made under the conditions of constantly rainy climates. Deposits of carbonaceous matter should be common.

The agent of transportation in regions of constantly rainy climates is almost entirely water. Floods are not entirely absent; they may be of rare occurrence. Waters are almost invariably more or less colored from contained colloidal organic matter, and only small loads of suspended inorganic matter are transported. The colloidal and dissolved loads in parts per million are usually small, but the total annual load of these substances may be large. There is essentially no atmospheric transportation.

Climate of Seasonal Dryness

Regions with climates of seasonal dryness have periods of rainy precipitation alternating with periods of little or no rain, each of variable length for the same region. Regions with this type of climate on the one hand may merge into those of semiaridity, as the eastern half of the Mississippi Valley merges into the semiarid region of the Great Plains, and on the other may merge into constantly rainy climates, as the same region passes into the wet lands of Louisiana and Mississippi. There may also be years or periods of several years when the climate is constantly rainy and others when it is semiarid with months each year when the precipitation is almost zero. Vegetation is correspondingly restricted, and the soils at these times become thoroughly dried and crack open to variable depths. Forest may cover the lowlands and also the uplands, but the uplands may have many open areas with scattered clumps of trees.

The general movement of the ground water is lateral and downward but may be locally upward during dry seasons, when rising waters evaporate at the surface and precipitate material from solution or colloidal suspension. These deposits frequently form concretionary structures, of which many are composed of the hydroxides of iron. Locally, these may become cemented together to form a firm crust that chiefly consists of hydroxides of iron together with more or less alumina and silica. This crust overlies the more or less insoluble constituents beneath. In most cases it is thought that the substances brought to the surface during dry seasons are washed downward and outward during wet seasons.

Decomposition extends down as far as, or a little below, the lowest level of the water table. There is a rather consistent leaching of carbonates, sulphates, and chlorides. The oxides and hydroxides of iron, hydrous aluminum silicates, and silica are left behind. The scarcity of vegetable matter does not favor reduction of the iron, and hence it remains in oxidized form. Under conditions of alkaline waters considerable silica is removed, and hydrous aluminum silicates may be reduced to silica and hydrous aluminum oxide. This reaches extreme development in tropical and subtropical latitudes, where the silica is largely removed and aluminum hydroxides are formed through reduction of hydrous aluminum silicates. The result is *laterite*, a residual deposit consisting largely of iron and aluminum hydroxides.

The colors developed in soils by climates of seasonal dryness range from gray and yellow to red. The yellow is developed under conditions of moderate or greater relief where the quantity of rainfall is such as to favor a fairly thick zone between the groundwater table and the surface and where the temperature is that of the polar halves of the temperate zones. The red develops under the same conditions of relief in tropical and subtropical regions. Relief must not be so great as to permit rapid removal of particles detached from parent rocks, for should such be the case, the colors might approximate those of the parent rock.

The rainfall of regions with seasonal dryness is spasmodic and may fall as torrents, at which times the quantity of water may be so great as to be beyond the absorptive capacity of the soils and vegetation, and there is much runoff. Floods are not uncommon. As a rule, all except the smaller streams flow throughout the year, but there are considerable differences in the streams of cultivated and uncultivated regions. Regions that had constant stream flow in the smaller streams when the region was primitive, after cultivation, may have no flowing waters in stream channels during dry seasons and may be in flood after every heavy rain. The sediments carried by streams consist of mixtures of inorganic and organic materials.

During the times of stream flood the waters may spread widely over floodplains. As the waters recede a part of the animal population of the streams remains behind in depressions. The waters in these depressions may later either partly or completely evaporate, and the animal populations are either partly or entirely killed and leave their skeletons either in or on the muds on the bottoms of the depressions. At times of flood the stream waters have a high content of suspended matter and are yellow to red therefrom. Considerable quantities of plant material ranging from macerated leaves to logs are also carried. The organic and inorganic sediments are deposited more or less together, reduction of iron compounds begins, and, if there is enough organic matter, colors become light blue to gray. If there is a deficiency, colors may be yellow to red. Deposition takes place upon alluvial fans about the heads of small streams, upon floodplains of rivers, upon deltas, and in the sea, and the deposits have characters determined by the environmental conditions. As reliefs of regions vary through a wide range and as there are places where the vegetation is inadequate to protect the products of rock destruction until decomposition is mature, the sediments removed may range from little-decomposed rock fragments to particles of mature decomposition. The larger rock fragments may be transported little or much, and when cemented are expressed as breccias, arkoses, graywackes, sandstones, and conglomerates. These may constitute large units in the rock sequence or be present as thin beds interstratified with claystones and siltstones. Dark shales and even coal beds may be occasional. Deposits made on lands of the climates considered are likely to become extensively mud-cracked as the dry seasons advance, and rain impressions should be common features.

FIRE, CLOSE PASTURAGE, AND CULTIVATION

It is considered likely that fires began with the advent of land vegetation, although it is thought that they were fewer before man appeared. On the other hand, it is thought that the fires before the advent of man may have covered greater areas and lasted for longer times than now, as they would have burned until extinguished by natural causes.

Fires may be placed in the three classes of forest, prairie, and swamp. Prairie fires temporarily eliminate the grass cover, but leave a dense mat of roots, so that not much in the way of erosion of soil materials is possible. Swamp fires may burn peat deposits for months to considerable depths, but except for destruction of the deposit little damage is done, and the impress on sediments seems negligible. Forest fires seem to have the greatest sedimentary impress. Not only may the trees be killed, but the vegetable litter to the top of the soil may be destroyed. This makes possible rapid removal of the soil cover on slopes, as the root mat is ordinarily not sufficiently great to prevent erosion. This removal would be expressed over sites of deposition in sediments passing from a certain degree of fineness, these sediments having been deposited when the plant cover was present, to some degree of coarseness, and finally to a considerable degree of coarseness as the materials at the base of the soil profile were reached, and the coarseness would continue until the deposit was built to the base level of deposition or until the plant cover was renewed. The change of texture from that present before the plant cover was removed to that after the fire had destroyed the cover indicates no diastrophism, no change of level, and no climatic change. It might be thought that the incident of fire would be shown by charcoal, but such would not be likely as the specific gravity of charcoal permits it to float and thus pass onward as the inorganic sediments are deposited. No proof has yet been presented that there are any parts of the geologic column for which fire is responsible (the burned coal beds of Montana and Wyoming are not considered), yet it can hardly be doubted that there are sedimentary features due to fire in parts of the column.

Close pasturage of grasslands may lead to complete destruction of the grass cover and ultimately to much erosion. The erosion provides draining streams with large loads which are deposited over alluvial fans, floodplains, deltas, and perhaps the sea bottom. Complete destruction of a grass cover might transform a grassland into an area of drifting sands, and these might encroach on adjacent forested areas. Complete destruction of grasslands through overgrazing is reported to have produced desert areas in South Africa (Bosazza, Adie, and Brenner). Whether there was close grazing before the keeping of herds cannot be stated, but it does not seem likely as there would generally have been sufficient predators to keep the grass-eating animals in check so that they would not exceed the sustaining capacity of the plants on which they lived.

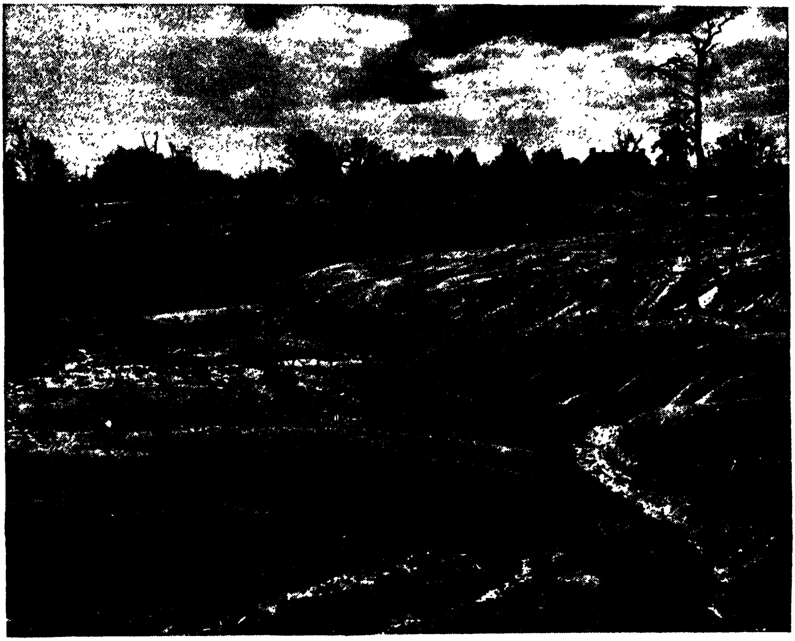

Fig. 4. This farm land near TVA's Pickwick Landing Dam in west Tennessee shows the effects of man-made erosion. The soil removed by the erosion may have been deposited over the Tennessee River floodplain and the Mississippi River floodplain, and some probably traveled to the Mississippi Delta and to the Gulf of Mexico. (*Soil Conservation Service.*)

Cultivation of land is a phenomenon of the past few thousand years, and it has been responsible for great increase in quantities of sediments transported to sites of deposition. In dry and semiarid regions it has destroyed the grass cover, pulverized the soil, and caused rapid removal of large quantities of soil by wind leading to the Dust Bowl of the Great Plains Region and the vast dust clouds of the past half century. In humid regions cultivation has removed the forest, destroyed the protecting litter of leaves and branches, eliminated the underground flora and fauna and the underground passages for which these organisms were responsible, and exposed the soils to the attack of falling water.

Fig. 5. Deposition due to human activities. Muds deposited over street and curb in Kenwood Park, Salina, Kansas, June 1938. Note the difference in the dimensions of the mud-cracked polygons over the street and over the parking. The mud probably has the same physical composition, but is thinner over the parking and probably contains more debris. (*Soil Conservation Service.*)

Ellison has shown the highly destructive effects of this exposure and Laws and Parsons state that the fall of 2 in. of rain on an acre has 6,000,000 ft.-lb. of kinetic energy. This vast quantity of energy applied to the soil loosens the fine materials, breaks up aggregates, produces muddy water, and leaves coarse materials. The muddy water infiltrates into the ground for only a very short period as very shortly passages for infiltration are closed through deposition of mud in suspension, making what Lowdermilk has termed the "colloidal seal." The water then flows away on the surface burdened to capacity with fine sediments. The results are great erosion, big loads to streams, floods, destruction of farm lands (Figs. 4, 5), and large deposits over sites of deposition. In dry land areas, the destruction of the native vegetation bares the naked soil to the destructive work of wind and vast quantities of wind-borne sediments are removed in a very short time (Fig. 6).

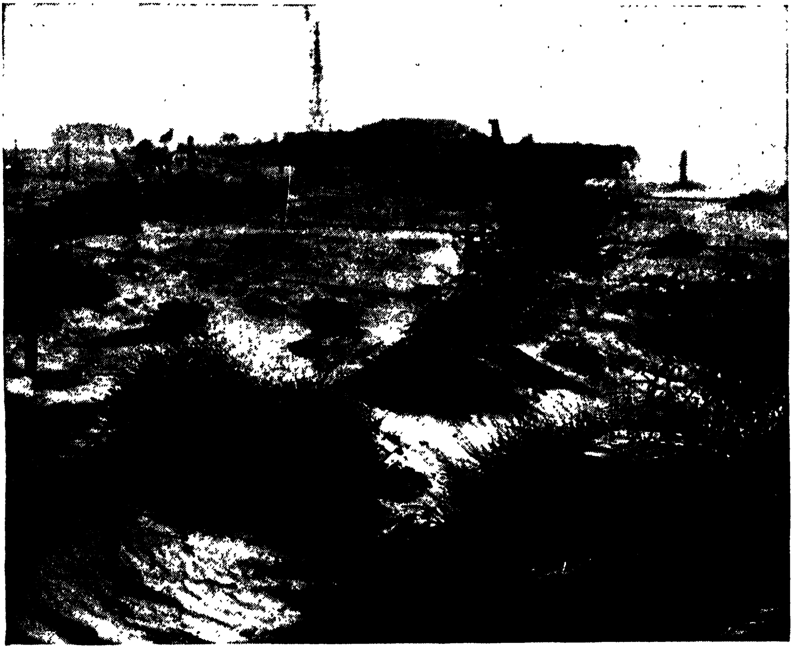

FIG. 6. A naked soil attacked by wind. This is man-made erosion, and sediments are transported to sites of deposition in a few hours; under natural conditions they would have required centuries for their removal. (*Photograph of an area in Texas by the Soil Conservation Service.*)

The effects of fire, pasturage, and cultivation are thus essentially the same. The results with respect to acquirement of sediments are greater the steeper the slopes subjected to these phenomena and approach zero as the slopes become gentle. The results from the depositional point of view seem greatest with respect to fire, as that has prevailed for a much longer time.

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CHAPTER III

CLASSIFICATION AND CONSIDERATION OF ENVIRONMENTS

INTRODUCTION

Any consideration of environments must face the problem of classification. This may be done on several bases, but it should be understood that any classification is arbitrary and that any environment passes by more or less gradual transitions into others. Likewise, as previously noted, little or nothing can take place in an environment that does not change it to some degree, and thus environments also vary in the time as well as in the space relationship. Lateral change in environments in many cases is extremely abrupt, and, similarly, environments may pass in time relationships into others as abruptly as, for instance, a rapid fall of sea level may change a sea bottom to land in a matter of a few hours as was done in the Yakutat earthquake in 1899.

As environments are responsible for the characters of the sediments that are deposited, a sedimentary response should be a sequel to every environmental change. This has been determined in some cases. As sedimentary consequences of environmental changes become better known, it may become possible to visualize from sediments the environmental factors responsible for them.

A most important basis for separation of environments is whether the medium that covers the surface is water or the atmosphere. Another important basis for division is whether the environment is on the continents or in the oceans, the islands being considered with the continents. This gives the two divisions of continental and marine. This seems more satisfactory than use of the overlying medium as a basis for division. The waters of the land are mostly fresh and are generally so different in their life, conditions, and sediments from the same things in the sea that alliance of the fresh water is closer with the land than with the sea.

There are places where continental and marine environments pass into each other. Bodies of inland waters like the eastern part of the Baltic Sea have neither a normal marine environment nor a normal fresh-water environment. There are lagoons behind barrier islands of which the waters range from marine to fresh. The tidal belt of the sea is exposed for longer or shorter periods twice each day. Estuaries have water replaced over parts of their areas with changes of the tides, and over other areas salt water flows upstream on the bottom while fresh water flows seaward on the top. Deltas have more or less intermingling of fresh and salt water or replacement of one water by the other.

The marine environment may be divided on the basis of depth, but there is no such thing as a natural division. Each of the subenvironments made on the basis of depth may in turn be subdivided on the basis of temperature, but again there is no natural basis of division, and the divisions made are applicable only to the surface and shallow waters, as the temperatures of the deep water are constantly so near freezing that divisions are not possible. The shallow-water marine environment may be further considered on the basis of stationary, rising, or falling sea level, as these conditions determine the thickness and characters of sediments that may be deposited at any place and the succession of organisms that inhabit the waters and the bottom.

TABLE 1. CLASSIFICATION OF ENVIRONMENTS

- Continental.
  - Terrestrial.
    - Desert.
    - Glacial.
  - Aqueous.
    - Fluvial.
      - Piedmont.
      - Valley-flat.
    - Paludal.
      - Lake swamps.
      - River swamps.
      - Flat-land swamps.
      - Paralic swamps.
    - Lacustrine.
      - Fresh.
      - Salt.
    - Spelean—cave.
- Mixed continental and marine.
  - Littoral.
  - Delta.
  - Marginal lagoon.
  - Estuary.
- Marine.
  - Neritic.
  - Bathyal.
  - Abyssal.

Classification of continental environments seems best made on the basis of whether or not the processes are aqueous. This gives the two divisions of terrestrial and aqueous.

The terrestrial division of continental environments may be divided on the basis of climate and physiography into desert, prairie, forest, high moor, and mountain, but, the desert excepted, the sediments deposited are due to aqueous conditions.

The aqueous environment may be placed in the divisions of surface and underground, and the surface waters into standing, flowing, or stagnant. The standing waters fall into the two divisions of fresh and salty.

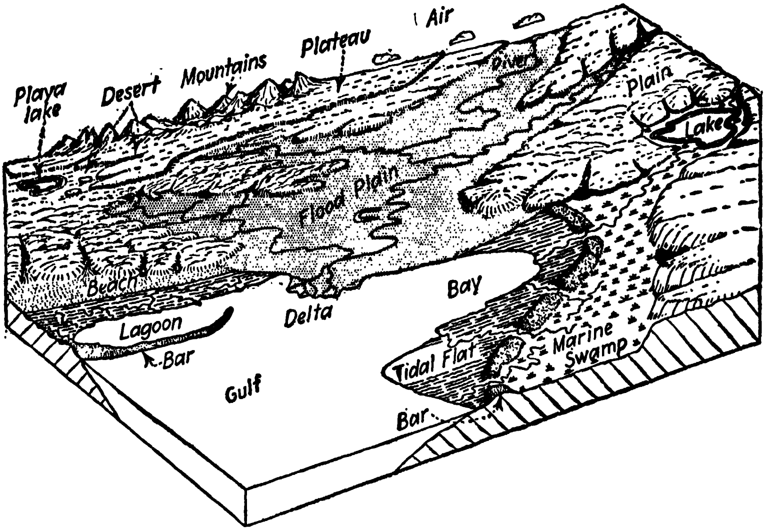

Fig. 7. Continental and marine environments. Alluvial fans are shown where several streams leave the highlands. (*After Twenhofel and Shrock.*)

The diagram of Fig. 7 attempts to depict these different environments, and they are also given in Table 1.

No exactness is claimed for the classification. Bases for division in many cases are arbitrary and very artificial, as illustrated by the separation of the neritic from the bathyal environment at the depth of 600 ft. Furthermore, any area over which a certain environment dominates may contain local areas where another environment exists as instanced by a lake or a swamp on the floodplain of a stream.

CONTINENTAL ENVIRONMENTS

As shown in the table, continental environments are placed in the two divisions of terrestrial and aqueous. Water, however, is almost invariably present to some degree in every terrestrial environment. It has primary importance in aqueous environments.

TERRESTRIAL ENVIRONMENTS

The Desert Environment

INTRODUCTION

About one-fifth of the present land surface is without drainage to the sea because of being too dry to permit streams to carve channels thereto. This approximates 11.5 million square miles. Not all of this is desert, and only small parts of this great area are receiving sediments. The largest existing desert, the Sahara, has an area of 3.5 million square miles, but sediments are being deposited over only about one-fifth of this area, or 700,000 square miles.

Sediments of deserts are of three origins. One is deposited by wind and consists of dusts and dune sands. A second consists of sediments deposited by currents of water brought into temporary existence at times of rainfall, or by permanent streams flowing from highland areas. A third consists of the sediments deposited in lakes that exist for longer or shorter periods of time in many deserts.

Deserts are due to four sets of conditions: small precipitation, infertility of soil, cold climate, and close pasturage and unwise cultivation. Only deserts due to small precipitation are considered in this book. Small precipitation may be due to an area's being in the rain shadow of a mountain chain or group, in the trade-wind belts, or on a warm land on the leeward side of cold waters.

In considering the desert environment, one in the rain shadow of a mountain is used, as in most respects other deserts due to dryness fit into the picture, as there are probably no large ones that do not have mountains in them and thus few that do not have sediments due to the presence of mountains.

A rain-shadow desert has three parts: (1) the mountain slopes, which may receive considerable rainfall and thus support plants; (2) the graded piedmont slopes and pediments, the former covered with debris from the mountains deposited by water and by slide and mudflow, the latter a rock floor with limited debris; and (3) a central area.

The mountain slopes and most parts of pediments are given over to erosion; materials upon them are those too large for the agents of transportation to move. The graded piedmont slopes are at times sites of deposition. The lowest part of the central area may contain a permanent water body which is salty and from which evaporites are precipitated, or it may be a dry lake or playa that contains water at times and at other times is dry, the surface then being either mud-cracked or dusty by reason of the crystallization of salts in the muds.

Deposition takes place on the graded piedmont slopes where and when waters flowing from highlands become loaded beyond capacity on slopes of lower inclination. At all places where conditions permit, winds are searching the surface and removing all materials within their competencies. The central area, if not covered by water, is attacked by wind, and all dusts and sands are removed, the former upward and perhaps completely out of the desert, the latter to some leeward position with respect to sources. The balance between deposition and erosion in the central area is determined by the permanence of water. Some desert basins that are wet for much of each year are filling because deposition exceeds depletion, whereas others that remain wet but a short time may be made deeper during each dry spell. In the former case the central-basin sediments advance upon the piedmont slopes; in the latter case the piedmont slopes extend farther into the desert basins.

DESERT FEATURES

Important desert features are the characters of the plants and animals dwelling there, the nature of the surface, the rainfall, and the methods by which rocks are destroyed.

Desert Plants and Animals. Deserts are not without plants and animals. The plants may be placed in four classes based on ability to make the best of the moisture situation. One class is composed of perennials and shrubs that have long taproots to enable them to go deep for water. A small plant a foot or so high may have roots many times its height. This class of plants lives where the water table is not so far from the surface as to preclude acquirement of water. A second class is composed of annuals and biennials. These live in deserts that have an annual moist period during which the plants quickly mature their seeds and die. Plants of this group usually have shallow, wide-spreading roots so as to acquire as much moisture as possible in a limited time. Plants with similar roots also dwell in swamps. A third class is almost without leaves and has changed its stems to succulent structures for storing water. Transpiration is reduced to a minimum. Here belong the various cacti. The fourth class includes many shrubs that use little water and have small leaves that minimize transpiration. Many of these plants are spiny, and among them are various sages, the creosote bush, and greasewood.

The indigenous animals of desert regions are not of wide variety, and the numbers per square mile are not great. All are adapted to rigorous conditions. There may be a large and unadapted temporary population when plants flourish and temporary lakes form during the moist season. As the vegetation dies and the food supply decreases, the migrant animals tend to throng the declining lakes, leaving their tracks and perhaps finally their bones on the surface to lie until the next rainy season, when they may be buried. The only other possibility of burial is beneath drifting sands. There are many chances for destruction of the bones, but the apparently less preservable tracks have better chances for preservation, as they cannot decompose or pass into solution, scavenger and predatory animals will not carry them away, and they can easily be buried. Examples in the geologic column that may record events of this kind are in the Newark Series of the Connecticut Valley, in the sandstones and shales of which tracks are locally abundant but bones are rare.

Surface of a Desert. The surface of a desert is due to both erosion and deposition. Erosion is done by water and wind, the former around the highlands of the region—the mountain slopes and the pediments—and the latter over the whole area. Depressions are cut in unexpected places and occur because the composition of the rock is such that it is easily cut by the wind or because there is concentration of the winds due to some combination of physiographic conditions. In the Desert of Gobi the lengths of these depressions range from 100 meters to 30 miles and the depths to about 100 meters (Berkey and Morris). The downward limit of aeolian erosion seems to be the level of the water table. The winds seek all places of physical weakness in the rocks and cut cavities that range from very small to large caves. These are bounded by rocks that may have undergone little decomposition or solution. The cutting is largely done by sands carried by winds, and as the sands tend to remain close to the ground, erosion is necessarily concentrated about the bases of cliffs, so that ultimately these overhang the bases. This erosion results in mesas, buttes, rocking stones, mushroom rocks, and etching of all exposed surfaces. The central area, if without a lake or playa, may be covered with dunes. If a lake or playa is present, the dunes may encircle this or lie around the leeward side. Dunes may also be present on the leeward side over high areas of a desert. Surfaces subject to active aeolian erosion are covered with lag gravels and boulders which are left because of the inability of the wind to move them. Many of these are faceted to form the well-known ventifacts or dreikanter.

Desert Rainfall. The outstanding feature of desert rainfall is the combination of extreme general scantiness and an occasional "cloudburst." A desert may bake in the sun for months, and then water may fall in torrents to form a stream in every depression leading from high to low places. This water more or less rapidly disappears in the thirsty sands. In some cases it spreads out widely over the low places to change a sun-baked, salt-encrusted playa of one hour into a broad, shallow lake of the next, which in the Black Rock Desert of northwestern Nevada has been known to cover 400 to 500 square miles (Russell). Such torrential waters do much erosion over unprotected surfaces and carry much sediment to central areas.

Rock Destruction. It has generally been assumed and stated that physical processes are mainly responsible for the rock destruction in deserts, that insolation is the chief agent, and that it is done in the form of block disintegration, mainly spalling and exfoliation, and mineral disintegration. Blackwelder (1925, 1933), however, considers chemical processes far more important than physical and does not think that changes of temperature are as important as generally supposed. It is thought that Blackwelder's view that chemical processes are of greater importance in desert rock destruction than physical applies largely to rocks susceptible to oxidation, carbonation, and hydration. It is also thought that these processes extend through a wider vertical range in deserts than in any other environment, with the result that most decomposable rocks have undergone some degree of decomposition long before they reach the surface. In deserts underlain by sedimentary rocks that are little affected by atmosphere and limited water, there would be little or no decomposition and certainly negligible solution, and all breaking of such rocks not due to earth movement would have to be assigned to the physical agents of rock destruction. Feldspars and ferromagnesian minerals of igneous rocks are rarely fresh for long distances beneath the surface. Blackwelder's studies in the California and Nevada deserts led to the conclusion that insolation is not important, stream corrosion has minor effect, wind abrasion is rarely conspicuous, frost action is locally important, and chemical destruction and the breaking of rocks by diastrophism and hydration are most important. However, there are places where wind abrasion, because of physiographic conditions, has considerable significance, and the dimming of automobile windows by desert sands is a measure of what may be done in a short time.

Desert Transportation. Transportation in desert regions is done by water, slide, mudflow, and wind. Strong currents of water, slide, and mudflow produce heterogeneous deposits about the upland areas, and water deposits stratified clays and silts in the lakes and playas. Rates of removal by wind are not known, but a minimum of 3.5 to 4 meters is known to have been eroded from Danby Playa in the Mohave Desert, as shown by small mesas capped with gypsum (Blackwelder, 1931). Many have emphasized the importance of deflation, and within recent years ocular demonstrations have been staged in the dust storms of the western plains (Fig. 6).

Desert Sediments. The deposits of the desert environment consist of poorly sorted and poorly stratified heterogeneous materials on the piedmont slopes; lag gravels, many of which are ventifacts, over the sites of active aeolian and aqueous erosion; sands marginal to, or on a side of, central areas and leeward over higher parts; and silts, clays, and evaporites over the lowest part of the central areas where these are not covered by sands.

Evaporites tend to be thin unless the deposits are made in bodies of water connected with the sea. Other deposits in the lakes may be iron carbonate and flint or chert. The gravels may have any composition; the sands are mainly quartz, but may be feldspar, calcite, or rarer substances. Minerals of considerable chemical instability are commonly present. Sand particles may have mat surfaces and both large and small grains may be well rounded. Much of the fine material seems to be powdered rock and not clay, but clay may be present to a greater or less extent.

The coarse deposits on the piedmont slopes may have considerable areal extent and locally may be thick. These deposits are fan-shaped, of aqueous and mudflow deposition, and have poor sorting and stratification. Boulders may range to large dimension, and mudflow seems to have provided the transportation for these. Lithic units are more or less crudely lenticular.

The Shinarump conglomerate of the Colorado Plateau seems to have originated in an environment about the uplands of a desert area. It is composed of lenses of gravel and overlapping lenses of coarse and fine sands. Sorting and stratification are generally poor. There is much cross-lamination with wide variation in the vertical and horizontal components. The Overton fanglomerate of Nevada is described as having similar characters and is interpreted by Longwell as having been formed in an arid country of high relief.

All fine sediments contain soluble materials and many contain calcareous concretions of various shapes, sizes, and exteriors.

The silts and sands deposited in water bodies of deserts may be much ripple-marked. Symmetrical or wave ripples are probably more common in small bodies of standing water than current ripples, with perhaps the ratio in reverse in large bodies. The stream and dune sands have current ripples only. Ripple marks of playas and dry lakes may be repeatedly destroyed by crystallization of salts in the deposits. Mud cracks and mud fragments made by breaking of mud-cracked polygons are likely to be common in all sediments in which mud cracks form. These have forms, spacings, and depths adapted to composing materials and conditions. Both aeolian and aqueous cross-lamination are common in desert sands.

Colors of desert sediments depend to a considerable degree upon the rigor of the aridity. In places of good underground drainage with few deposits of salts in surface materials there may be considerable oxidation, particularly in tropical regions, and yellows, reds, and browns are developed. Colors of piedmont sediments seem more likely to be determined by the original colors of the composing materials. Deserts of very rigorous aridity seem to be characterized by light-colored deposits unless the country rocks have other colors. Drifting sands are light-colored in the range from white to yellow. Clays and silts of the lakes and playas range in color from white or gray to black. Black is due to carbonaceous matter or the black monosulphide of iron produced from bacterial reduction of sulphates. Black due to carbonaceous matter will persist, but where due to the monosulphide of iron it will fade as the monosulphide changes to marcasite or pyrite. Evaporites of the lakes are ordinarily white. Over areas of rising waters due to evaporation the colors are largely due to the salts precipitated in the surface materials, and light colors dominate.

After deposition there may be introduction of substances with consequent change of color. If the introduced materials are iron salts, the sediments may become yellow to red. This is believed to have taken place in the sands of the Nefud Desert in Arabia described by Phillips, where sand hills, seemingly dead dunes, have the sands covered with films of iron oxide.

The picture thus presented of the deposits of the desert environment shows an interlensing of various kinds of clastic deposits made by water and wind that range from very coarse over the piedmont slopes to very fine in the lakes and playas. Bedding of the piedmont deposits tends to be very poor and sorting is equally so. Bedding and sorting of the lake deposits tend to be excellent and may be equally so in the playa deposits. Sands of dunes should have excellent sorting. The sands, clays, and silts of the lakes and playas may have interbedded evaporites. Sediments over lower parts of deserts may contain considerable calcium carbonate and other salts. Ventifacts should be present among lag gravels, and many rock fragments should be coated with desert varnish. Plant and animal remains should be rare, but tracks may locally be common. Primary colors of sediments tend to be light, but reds and yellows may be present, and some lake sediments may be black.

The thickness of sediments that may accumulate in the desert environment is not known, but 300 meters is considered a fair maximum for the entire association. Conditions may be pictured under which a greater thickness could accumulate.

Before sediments may positively be referred to origin in the desert environment the entire association described above should be present. It is known that many of the features described may form under other conditions. Wind-blown sands are deposited under many conditions. Ventifacts may be made wherever winds blow; desert varnish is widely present in regions that are not deserts, and a somewhat similar feature may form on boulders in the tidal zone of some shores; and evaporites may form under conditions that have nothing to do with deserts. Moreover, sediments may originate in the desert environment and be deposited in another. There has been too much assignment of sediments to a desert origin on the basis of only a few features.

EXTENT OF DESERT DEPOSITS

Deposits of deserts are of wide extent at the present time on all continents. An approximately equal extent may be assumed to have obtained during most of the geologic periods. If current conceptions of absence of land vegetation during the Pre-Cambrian are correct, there should be extensive aeolian deposits in the rocks of Pre-Cambrian systems, even if there was no aridity. The oldest sediments that have been ascribed to arid conditions are in the Torridonian sandstones of Scotland and the Eophyton sandstones of Sweden, from which ventifacts have been collected, but there does not seem to be much in the deposits permitting interpretation of aridity. The Upper Cambrian sandstones of the upper Mississippi Valley contain sands that seem to be of aeolian production, but there is no evidence of aridity, and in most cases deposition was aqueous. The St. Peter sandstone of the Mississippi Valley has been referred by some geologists to desert origin. The sands carry evidence of aeolian production, but the chief agent of deposition was certainly water. The Sylvania sandstone of northwestern Ohio and adjacent Michigan was ascribed to deposition in the desert environment by Grabau and Shimer on the basis of the clean, well-sorted, and well-rounded sands, white color, and stratification and cross-lamination, but Carman has shown that in northwestern Ohio most of these sands are of aqueous deposition.

The Permian and late Pennsylvanian witnessed arid and semiarid conditions over large areas in north central Europe and central western United States, and the complete association of deposits of desert environments seems to be present in central Europe and some of the areas of the United States. The sands are red in both areas, and this color seems to be largely primary. The associated evaporites seem to have been deposited largely in bodies of water connected with the sea. It is assumed that the red sediments were developed in warm and humid upland regions and washed into waters of high salinity or upon dry lowland areas bordering the sea. Redness of color was retained because absence of organic material precluded reduction of the iron. Some Permian "Red Beds" of western United States seem to have been deposited under dry but hardly severe arid conditions.

Triassic deposits of England seem to offer the complete association of desert deposits in the presence of wedge-shaped units with aeolian cross-lamination dovetailing with stream-deposited gravels and sands marginal to highland areas, lenticular nature and lack of persistence of units, and beds of clay containing layers of gypsum and rock salt (Lomas). Triassic strata of western United States have been referred to deposition in the arid environment, and there seems little doubt that some of them are properly so assigned.

On the Colorado Plateau is an extensive sandstone formation known under the names of La Plata and White Cliff. This has been interpreted as having been deposited in a desert that lay south of the Jurassic Sea. The sandstones have a maximum thickness of 600 meters, and the color is generally white, but locally it may be red. Some beds are over 20 meters thick and do not have an internal trace of bedding. Others are highly cross-laminated with the units wedge-shaped, as is characteristic of aeolian deposition. Other beds have aqueous cross-lamination. The complete association of desert deposits is not known to be present, but it seems certain that aeolian deposition took place over large areas.

Parts of the Cretaceous Cheyenne sandstone of Kansas were deposited by wind, and Bailey has suggested that the lands bordering the Cretaceous seas of France and England in which the chalk was deposited were hot deserts like the Sahara because of the presence of well-worn grains of quartz in the chalk. The reason is a poor one, as grains of like character may be found in many calcium and magnesium carbonate deposits. There is no evidence that conditions were arid.

There may be other deposits of the desert environment in the geologic column that have been referred to other origins, because of the traditional view that nearly all sediments are aqueous. Differentiation is needed, however, between aeolian deposition and aeolian production, neither of which denotes a desert environment, and deposits of arid conditions.

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The Glacial Environment

The glacial environment at the present time covers most of Greenland, the whole of Antarctica, and large areas in high mountains; in some of the past geologic periods vast areas of the earth's surface were the domain of glaciers. At the present time over 60,000 square miles of Alaska, an area greater than *that of Wisconsin or New England, is covered with glaciers or is receiving glacial or fluvioglacial sediments.*

The environment of the glacier is characterized by low temperature and limited plant growth. The sediments are deposited from ice or from ice-cold waters. Glaciers are margined about their ends by end or frontal moraines which are partly of ice and partly of melt-water deposition. These may have kame deposits on the stoss side, which are made where streams flowing from glaciers have their velocities checked by the obstruction of a moraine. Small temporary bodies of water may lie between the kames, held by the ice and the moraine, and standing-water deposits are made in these. A glacier may dam a valley and create a lake. Under these circumstances the end moraine is buried beneath melt-water sediments deposited as small deltas, known as *delta kames*, where streams leave the glacier. The waters that form kames must ultimately cross the moraines, which is done in one or several places, and thus melt-water deposits may extend from the stoss to the lee side. On the lee side there is usually no obstructing moraine, and the usually heavily loaded waters rapidly aggrade the surface to form outwash deposits.

As a glacier retreats it leaves both ice and melt-water deposits over the area deserted, the melt-water deposits usually above. At each halting place in the retreat a new end or frontal moraine is built. Each moraine may serve as a dam, and a part or even the whole of the area between two moraines may be covered with standing water in which melt-water deposits are made. The irregular deposition of glacial materials forms many small to large depressions. Melt-water deposits are also made in these. Floating ice is probable in these bodies of water, and if these drag bottom, stratification is disturbed. Winds may cover some glacial and melt-water deposits with dusts derived from the surface of a glacier and from glacial and melt-water deposits. Advances and retreats of glaciers may superpose a succession of deposits of the glacial environment.

Materials deposited by ice are unstratified, unsorted, and composed of a wide range of material, in both composition and dimension. The melt-water deposits on the stoss side of a moraine are laid down by water carrying an excess of sediments and rapidly losing velocity, with the consequence that high initial inclinations may form, sands become steeply cross-laminated, and the sediments vary greatly in dimensions of particles in successive layers or laminations. On the leeward side of a moraine the waters do not lose velocity so rapidly, with the result that sediments are better sorted and stratified, and the cross-laminations are not so steeply inclined. Standing fresh-water deposits may be varved and consist of clays and silts. The summer part is lighter in color, coarser grained, and thicker than the winter part. The summer part is composed of some mixture of sand, clay, and silt; the winter part is very fine clay. The thickness of the summer part is more or less proportional to the coarseness of the composing materials. The winter part tends to maintain an approximate equality of thickness. Summer parts pass gradually into those of winter; a winter part and the overlying summer part are rather sharply defined from each other. If melt waters flow into salt lakes or the sea, the fine sediments are flocculated and varves do not form.

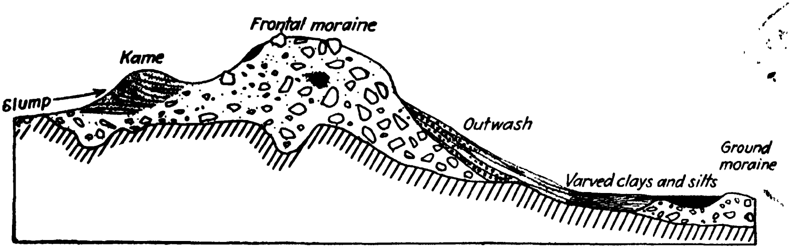

FIG. 8. Cross section of a glacial deposit showing the association of ice and melt-water deposits.

Glacial and melt-water deposits are rather generally without fossils because of low temperature. Colors usually range from white to gray-blue, but intrinsic colors of minerals may give other colors. Occasional rock particles may be striated, and the surface over which ice moves may be both furrowed and striated.

Deposits of the glacial environment contain decomposed materials only to the extent that the sources of the materials were decomposed. Ordinarily they are small to large pieces of the original rocks and minerals.

Deposits of the glacial environment usually do not exceed 100 meters in thickness, but some mountain glaciers have made deposits 300 or more meters thick.

Proof that deposits were made in the glacial environment requires that the entire association of deposits, from those of the stoss side of the moraine to the varved clays of the outwash lakes, be present, and the case is strengthened if the surfaces on which the deposits rest carry marks of glacial movement. Lack of stratification and poor sorting prove little, and even the presence of striated rock particles is not conclusive. The associations shown in Fig. 8 represent proof.

EXTENT OF GLACIAL DEPOSITS

The oldest glacial deposits in the geologic column of North America are in the Lower Huronian of Canada, where they cover an area with an east-west extent of 1,000 miles. Other Proterozoic glacial deposits are the Numees tillites and its equivalents of South Africa, deposits in the Wasatch Range of Utah, and in South Australia the Sturtian tillite has an extent in a north-south direction for 300 miles. Slates in this formation, interpreted as varves, are 300 meters thick. Other Pre-Cambrian rocks interpreted as tillites are present in Norway, east Greenland, China, and India. Some of these have been referred to the early Paleozoic.

Glacial deposits are extensively known from the late Pennsylvanian and Permian. These have extensive distribution in South America, South Africa, India, and Australia. The South American occurrences are in Argentina and southern Brazil and form the Santa Catharina formation. In South Africa they form the Dwika conglomerate with a thickness of 300 meters, an east-west distribution of 600 miles and with movement for 500 miles south from a source near the southern border of the tropics. In India the Talchir conglomerate represents deposits of the glacial environment. The movement was from south to north. The only glacial deposit in North America assignable to either the Pennsylvanian or Permian period is the Squantum tillite of Massachusetts and Rhode Island. This deposit is thought to have been made by a mountain glacier.

Cretaceous tillites have been described from central Australia, where they are said to cover 40,000 square miles, and tillites of Eocene age have been described from Colorado and British Columbia.

The recent Pleistocene glaciation placed glacial deposits over several million square miles on the lands around the North

Atlantic to form one of the greatest of sedimentary formations.

According to Flint nearly 6 million square miles of the present land surface is now covered with glaciers and considerable parts of this and some bordering areas are receiving glacial deposits. This is over 10 per cent of the existing land areas.

Aqueous Environments

Fluvial Environments

Fluvial environments are divisible into the headwater part of the stream, designated the *piedmont;* the main part of the stream, termed the *valley-flat;* and the *delta.* As the delta is built into standing water and thus has a dual character, it is considered a transitional environment and given separate consideration. There are no sharp distinctions separating these three environments, as each passes invisibly into the other. Each is typified by certain associations. In each phase of fluvial environments may be found examples of paludal, lacustrine, and aeolian deposition.

THE PIEDMONT ENVIRONMENT

The piedmont environment is found over that part of the courses of streams where steep gradients of upland regions pass into the gentler gradients of bordering lowlands. The environment passes gradually into that of the valley-flat. Physiographically, the deposits made in this environment form the well-known alluvial cones and fans and talus slopes. Deposits are largely made by streams, but creep, mudflow, slide, and rockfall may make important contributions. The relative contributions from these various sources vary with conditions, among which temperature, rainfall, slope, and character of rocks have importance. Great extents are favored by steep slopes, weak rock, much rainfall over upland areas, glaciers at the heads of streams, much freezing and thawing in the uplands, subsidence of the areas of deposition, great extent of upland area, semiaridity over piedmont areas, and renewal of elevation from time to time. The surface of a deposit slopes away from the region of supply of sediments and also away from the stream of deposition.

Inclinations of stratification units are high on the upper parts of cones and fans and may approximate the angles of repose for the materials. The slopes over the main parts of fans range to 15 or more degrees and decrease toward the base finally to pass into the lower inclinations of the valley-flat.

Streams flowing across growing piedmont deposits gradually fill and dam the channels in which they flow. These are then abandoned and new channels are sought between the old. Channels are invariably shallow. Piedmont deposits in process of degradation have the streams in channels that tend to be deep and narrow (Fig. 9).

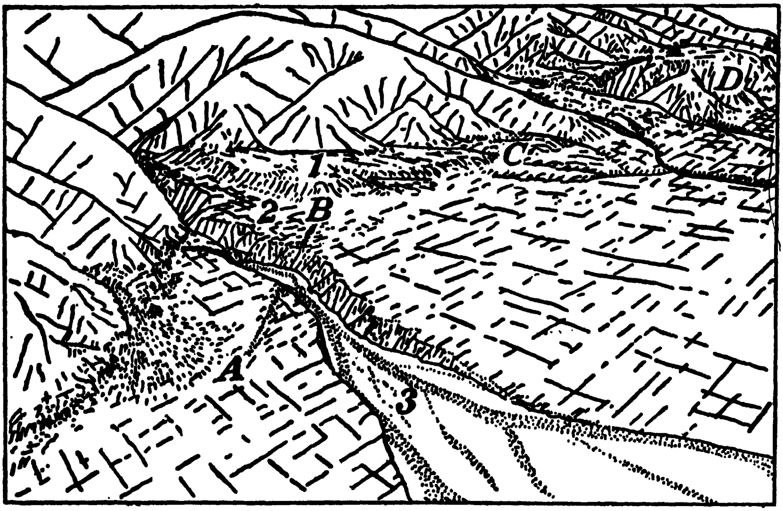

FIG. 9. The mouth of Cucamonga Canyon from the southwest and the Cucamonga and associated fans. About 1½ miles of mountain front is shown. *ABCD*, fault scarp. 1,2,3, alluvial surface. (*After R. Eckis, Alluvial fans of the Cucamonga District, southern California, J. Geol.*, 36 (1928), 229.)

As the deposits of the piedmont environment are composed of more or less porous materials, much water is lost by absorption as streams progress across them. The streams may disappear from this cause and thus deposit their entire suspended and tractional loads. Some of the water thus absorbed may reappear at the lower ends of piedmont deposits.

Piedmont deposits range from clays and silts to blocks of large dimension. Piedmont streams are known to have carried blocks with weights to 17 tons, and larger blocks may be introduced by mudflows and slides. Rounding is not likely to attain great perfection. When lithified, the deposits form a *fanglomerate*

(Lawson) composed of a more or less heterogeneous association of units of conglomerate, sandstone, arkose, graywacke, siltstone, and claystone, the proportions of each varying with locality. The upper part of a piedmont deposit is generally composed of the coarsest materials transported—gravels and probably boulders. Sands generally dominate over middle areas, and sands and clays are the chief sediments over the lower areas of a piedmont. Fossils are more often absent than present, but there may be local occurrences of abundant fossils, due to sudden floods which caused wholesale killing and rapid burial.

Colors vary somewhat with climate and with sources of sediments, and colors of original rocks play a large role. Gray and yellow seem to be the colors most common, red and brown are occasional, and, if organic matter is present, the color may be dark to black.

Sorting and stratification are generally poor, or at least very imperfect. There are none in talus and mudflow deposits. Units are lenses that persist for but short distances, and the arrangement is perhaps best described as an interlensing of units of limited area. All variations of clastic sediments dovetail with each other. There is much cut-and-fill structure. Initial inclinations tend to be high and to approximate the inclinations of the surfaces on which deposition takes place. Sands and the finer gravels are likely to be cross-laminated with much variation in degree and direction of inclinations. Current ripple mark may be present, and mud cracking is occasional.

It is possible that deposition may form basins between the fans. Some may be deep enough to contain water in which clays, silts, lime carbonate, and perhaps evaporites may be deposited. In other cases the depressions may become swamps filled with plant accumulations. If piedmont deposits are made in dry regions, aeolian deposits may be incorporated with the aqueous.

A piedmont deposit may attain great thickness if the upland is in process of elevation so as to produce a continuous and not a declining supply of sediments, and the region of deposition in process of subsidence so as periodically to raise the level of the base level of deposition. Any thickness is then possible. If regions of supply and deposition are stable, the thickness is determined by the extent of the original elevation, which in turn determines the elevation of the base level of deposition.

The Piedmont Cycle. The piedmont cycle has a period of building and a period of removal. A piedmont deposit begins to form at a foot of a highland at some stage in its elevation. Continued growth leads not only to greater height and thickness, but also to extensions up the slopes of the elevated area and outward over the bordering lower land. This continues until the base level of deposition is attained. There follows a shorter or longer time of temporary deposition and removal over the surface of a piedmont area, after which permanent removal begins. The outer margin of the deposit then retreats sourceward irregularly just as it was built, and at the same time the upper margin retreats down the upland. In time the piedmont deposit is entirely removed, and the cycle is complete. It cannot be doubted that such cycles have been repeatedly completed about the ancient mountains of the past whose roots are now the only evidence of their onetime presence.

Existing Deposits of the Piedmont Environment. The earliest sediments that may be referred to formation in the piedmont environment are some of the coarse deposits in the Huronian, some of the Keweenawan conglomerates and sandstones of the Lake Superior Region, the Doré series of western Ontario, and parts of the Great Smoky and Cochran conglomerates of the southern Appalachians (Barrell). Parts of the Old Red Sandstone of Great Britain may be of this origin, some of the Pennsylvanian sediments of the Appalachians, and the New Glasgow conglomerate (Permian) of Nova Scotia. The Triassic Newark series of the Piedmont Region from Nova Scotia to North Carolina has piedmont deposits in most of the fault basins in which these sediments accumulated. Parts of the Flysch and Molasse of the Alps certainly accumulated as piedmont deposits, and such is also the origin of parts of the Siwalics of northern India. Tertiary piedmont deposits have extensive distribution over western United States, and to this origin may be referred most of the extensive clastic deposits that flank the Rocky Mountains on the east. The Sespe formation of California has been referred to this origin by Kew, Reinhart, and others and also the Overton fanglomerate of Nevada by Longwell. Piedmont deposits in process of formation may be seen on the north side of the Los Angeles basin (Eckis) and along the east front of the Sierra Nevada. They have extensive development in Argentina along the east front of the Andes (Barrell). Deposits of piedmont origin seem to be among the greatest on the continents.

THE ENVIRONMENT OF THE VALLEY-FLAT

Introduction. The valley-flat environment includes the floodplain and stream channel. It differs from the piedmont in greater permanence of stream position, so that differentiation into floodplain and channel is possible. The distinction holds fairly well except under conditions of excessive aggradation when channels are periodically filled and compelled to change position. The environment also differs from that of the piedmont in a lesser slope to its surface, far lower velocity of water, and in lesser capacity and competency of the water. Sediments transported are smaller and may be rounder. Sorting is also much better. Decomposition of the materials transported is generally more advanced. Methods of deposition are different in that there are no mudflows except possibly on the borders of the valley-flat. The valley-flat deposits pass by inappreciable gradations into the deposits of the piedmont headward and downstream into the deposits of the delta. Streams without floodplains are not considered, as these are cutting channels, and any deposits made are temporary and have little or no chance of preservation.

A stream channel consists of a succession of deeps and shallows in which the sedimentary processes are diverse and different at different times. These are considered under stream transportation. Deposition in a channel or the building of a delta ultimately compels deposition upstream. Decrease in supply of sediment leads to cutting in the channel, provided the base level of erosion has not been reached. Fall of sea level has the same effect. Deposits on the valley-flat under conditions of stationary crust reach their greatest extent both in area covered and in quantity of sediments deposited during that part of the land cycle of erosion from late maturity to old age.

The processes of deposition in the valley-flat environment are largely physical, but mingling of colloids and solubles from different tributaries is doubtless responsible for some deposition. Thus, some deposition should take place where the waters from the western plains of the United States with high content of dissolved sulphates empty into the Mississippi. Also there is some deposition due to organisms in most rivers, and chemical and organic deposition takes place in some floodplain lakes and swamps.

Many floodplains have large populations of plants and animals, and remains of these should be common in channel and floodplain sediments. Conditions on floodplains, however, are not always conducive to preservation, as rates of deposition may be slow and burial may take a long time. Animals and plants that dwell in floodplain lakes are likely to be preserved in fossil form. Occasional floodplain animals become mired in swamps and drowned in lakes, and the occasional flood may kill large numbers of animals in a short time and bury the carcasses in the deposits of the channels.

Valley-flat Sediments (Fisk). Streams make deposits on floodplains when they are in flood. Deposits may be made in floodplain lakes and swamps at all times. The floodplains of dry regions and to some extent those of humid regions may receive aeolian deposits in the forms of dust and sand, both being acquired from the stream deposits.

Floodplain deposits depend upon the sediments supplied and the local climatic conditions. Sediments supplied depend upon terrane, topography, climatic conditions of the places of origin, and the transportation. If climatic conditions permit a floodplain to maintain a plant cover, organic materials are mingled with the sediments. This reduces iron oxides, eliminates yellow to red colors, and produces gray to black colors. Floodplains of dry climates do not maintain a good vegetable cover, and the sediments tend to retain the colors of deposition; and if these are in the range from yellow to red, they may be expected to remain so.

Bedding may be wanting in floodplain deposits, particularly if deposition is slow enough to give ground organisms sufficient time to work each new deposit into the old. Mud cracking, prevalent over dry floodplains, may obliterate much bedding.

The sediments deposited on flood plains are ordinarily in the range from clay to sand. Gravels may be occasional, but the particles tend to be small. There may be considerable organic matter, and this attains greatest importance in floodplain swamps. Sediments may contain considerable soluble matter left by the waters that filled them when they were deposited. This may be concentrated on dry floodplains by rising underground waters to form concretionary structures like the *kunkur* of India, or *caliche*. Floodplain lakes may produce any form of lacustrine deposits. Dune sands may locally be present.

Sediments of channels are of physical deposition in the range from clay to gravel. There may be carbonate deposits locally made from shells of organisms, but it does not seem that these attain much importance. The sediments have extremely erratic distribution, which arises from the nature of stream action in the channels. There are many local unconformities, of which some have striking relief. Every sedimentary unit is a lens with great variation in thickness and horizontal extent. Sands are more or less cross-laminated, mostly downstream, but occasionally upstream. Current ripple mark is common with orientation related to direction and velocity of current. Wave ripple mark should be rare or wanting. There is much cut-and-fill structure (Fig. 10).

Sediments of the valley-flat environment are usually not very thick, but extensive deposits may form over floodplains of dry regions if the streams are provided with an abundance of sediments, and also if the environment is in a subsiding region, under which conditions the thickness is limited only by the extent of

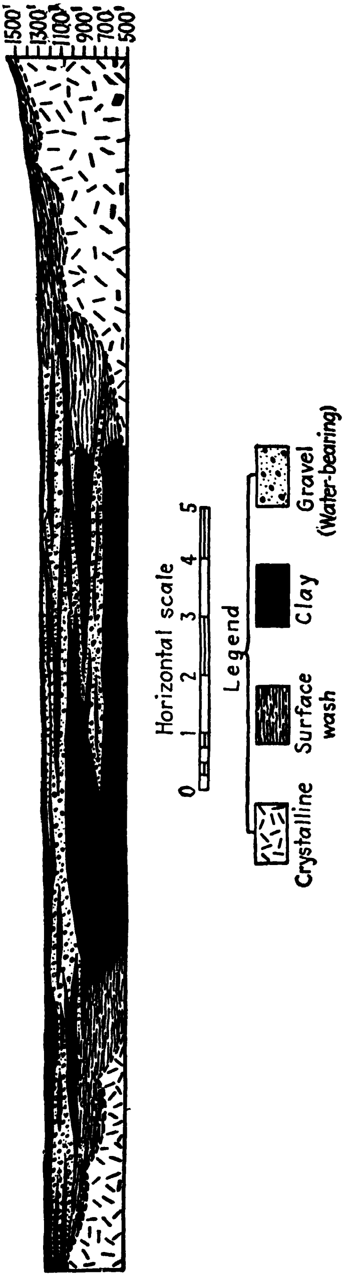

FIG. 10. Profile and cross section of a valley-flat deposit, Salt River, Arizona. The gravel deposits may be considered to represent the migrations of the channel as the deposits accumulated. (*After W. T. Lee, U. S. Geol. Surv., Water Supply Paper* 136 (1905), *Pl.* 21 *C.*)

subsidence. The rate of subsidence should not exceed the rate of building of the deposit, as such a condition might introduce a standing body of water and another environment.

Extent of Deposits of the Valley-flat Environment. Valley-flat deposits are now of extremely great importance over such great river valleys as the La Plata, Amazon, Orinoco, Nile, Po, Indus, Ganges, Mississippi (Fisk), and others. The areas covered attain many hundreds of thousands of square miles. The oldest valley-flat deposits are present in the Pre-Cambrian, of which parts of the Keweenawan of the Lake Superior Region and the Ocoee and Chilhowee Series of the southern Appalachians have characters that refer these parts to this origin (Barrell). The Catskill formation of the Devonian of North America and the Old Red Sandstone of Great Britain are partly of valley-flat origin. So are large parts of the Pennsylvanian of North America. Much of the Permian of western United States seems to have been deposited by streams. The Morrison formation (Jurassic) of the Rocky Mountain Region is composed of floodplain clays and sands bordering channels filled with sands and gravels. Some of the channel deposits contain many bones. The Dakota sandstone of the Great Plains seems to have been deposited by streams flowing from the east, and where this has been carefully studied, as in Kansas, it has been differentiated into the flood plain and channel deposits (Rubey and Bass). The Tertiary sediments of the Great Plains, of which the mountain parts are piedmont, are among the best examples of valley-flat sediments in the geologic column of North America (Davis, 1900; Barrell, 1925). These were first interpreted as lake deposits but are now rather generally assigned to stream deposition with some floodplain lake deposits. They consist of clays, silts, sands, and gravels. The clays and silts have color ranging from pale blue and pale green to gray and yellow, and the sands and gravels are gray to reddish. Most sediments contain considerable lime, which in some places has so cemented the gravels as to form the so-called "mortar beds" of Nebraska and Kansas. Beds of algal limestone are present in Nebraska. These probably represent deposits made in floodplain lakes. Remains of land animals are abundant in some beds, and locally there is much lignite. The great Siwalic formations on the flanks of the Himalayas in northern India with a thickness to 6,000 meters were in part deposited in the valley-flat environment (Weller). More or less recent valley-flat deposits are present in the Great Valley of California, the Puget Sound depression, and the Indo-Gangetic plain.

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The Paludal (Swamp) Environment

INTRODUCTION

Swamps develop in depressions that are not deep enough to form lakes, about the shallow margins of bodies of water that are invaded by plant growth, and under climatic conditions of high relative humidity and much and often-repeated rainfall they cover almost any surface. Many swamps support a growth of various reeds and grasses; others are covered with alders, or large trees, as the mangrove and cypress of the south and the tamarack of the north. Many extensive swamps are covered with moss.

Swamps cover large areas of the present surface of the earth, and the area in the United States in the swamp environment has been estimated as over 100,000 square miles. There are large areas in Canada and Alaska. Europe has extensive swamps in Ireland, Russia, the Scandinavian countries, and north Germany. Large areas of Siberia are swampland. Extensive swamp areas are in tropical and subtropical countries, as the great swamps of Sumatra, Florida, Virginia, the Carolinas, Mississippi, Louisiana, Central America, and elsewhere. It is probable that more than 1 million square miles of the earth's surface are covered with swamps.

The swamp environment has its most extensive development, if climatic conditions are favorable, upon plains of construction, as floodplains and deltas, and elevated sea bottoms. If climatic conditions are favorable, swamps may be coextensive with almost any surface. Swamps are classified as follows:

- Marine (paralic) swamps.
  - Grass-and-reed swamps.
  - Mangrove swamps.
- Fresh-water swamps.
  - Swamps connected with basins.
    - Lake swamps.
    - River swamps.
  - Swamps on flat or gently sloping surfaces.

MARINE (PARALIC) SWAMPS

Marine swamps may begin their history as a protected part of shore areas of the sea which may have been produced by a barrier beach, spit, or bar, by elevation of the sea bottom to make the water shallow for a long distance seaward, or by submergence of a flat-land area to produce the same results. The development of marine swamps and marshes is favored by small or no tides. Large tides create large waves and increase the difficulty of plant growth.

Grass-and-reed Swamps. The development of a marine swamp of the grass-and-reed type is perhaps best illustrated by following the history of such a swamp from a lagoon behind a barrier (Davis). Before the building of a barrier the site of the future swamp probably received shallow-water sediments. After the building of the barrier, the protected waters were less agitated, and, if sufficiently shallow, aquatic plants covered the bottom, among which eelgrass probably was among the first. Later, after some shallowing, such reeds as *Spartina* succeeded the first plants. These and other plants strained the passing water of sediments, and ultimately the bottom was built to the surface of the water. The sediments were filled with water and they probably settled as other sediments accumulated above them, so that plants and sediments originally grown and deposited at one level, in course of time, settled to a deeper one (Fig. 11).

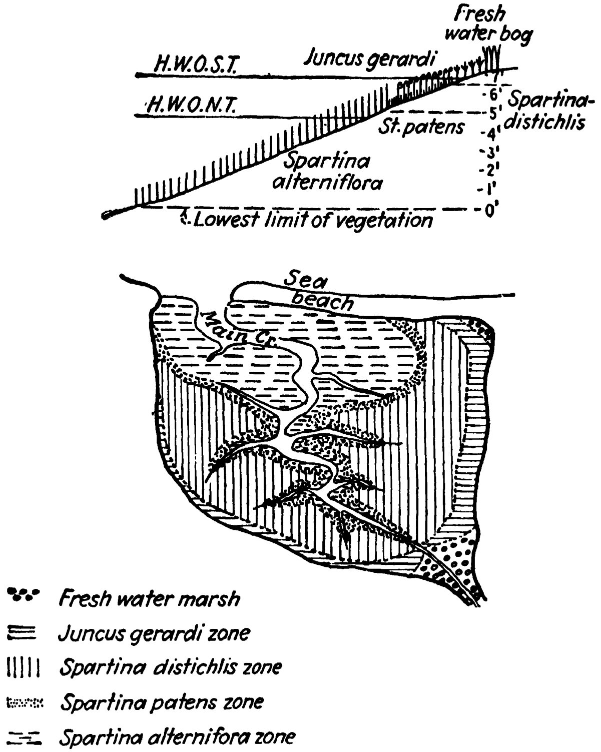

FIG. 11. Diagram showing vertical and horizontal distribution of plant zones in a New England marsh. The horizontal lines in the upper diagram show tide levels at spring and neap tides. (*After V. J. Chapman, Proc. Geol. Assoc.*, 49 (1938), 376.)

In this consideration a barrier has been assumed, but if the water is shallow for a long distance from the shore, this need not be present. On the Gulf of St. Lawrence between Harbour St. Pierre and Mingan a swamp borders the shore for several miles, and, on the east shores of the Baltic, marine swamps may be found in nearly every bay. The former has a tide of some 6 ft.; the latter is almost tideless.

The possible thickness of accumulation of sediments in marine swamps under stationary sea level cannot be great. Aquatic plants commonly do not begin to grow in waters more than 3 to 4 meters deep and they can build deposits only about a meter above sea level, so that a maximum thickness of deposits would be somewhat less than 6 meters. However, if a region of swamp accumulations is one of slow subsidence, a large thickness is possible, but the rate of subsidence must be nicely balanced to the rate of accumulation of sediments for this to be possible. If the rate of subsidence exceeds the rate of accumulation, the sea would cover the swamp with marine sediments.

Extent of marine-swamp deposits depends upon the extent of area protected. If the protection was a bar or barrier beach, it seems likely that the area would be small, but if the protection was due to shallow water and limited tides, a large area would be possible of which the outer part would be bathed by salt water and have a population of marine animals and the inner part might contain fresh water with adapted animals. Salt waters would circulate through the outer plant growth, and the marginal plants would be attacked by waves which would be small if the water beyond the plant growth was very shallow.

As a marine swamp develops, the sediments pass from completely marine to paludal. The lowest sediments are clays, sands, and calcareous materials. After the reeds and grasses make appearance and take possession, organic matter forms an increasingly larger proportion of the sediments and ultimately makes the whole of them (Steers and Thomas). Black shales should be common in early stages of a marine swamp, and if subsidence maintained the swamp at that level, they might form the main and culminating sediments and be filled with planktonic organisms floated in from the sea. It is suggested that some of the graptolite shales were formed in this way.

Waters of marine swamps contain a large content of sulphates.

These invite entrance of sulphate-reducing bacteria. Hydrogen sulphide is an ultimate product of the activity of these bacteria, and some oxidizes to sulphuric acid. This acts on carbonates in solution to form more sulphates. Some hydrogen sulphide leads to the formation of iron sulphide, which ultimately appears in the sediments as marcasite and pyrite.

Mangrove Swamps. Mangrove swamps are confined to tropical and subtropical regions. The trees must have their roots bathed in salt water, and they extend themselves seaward over all bottoms sufficiently shallow for them to grow. Their interlocking roots form dense mats, which catch and hold sediments, and ultimately a firm and compact substratum forms. The mangroves then die, and other forms of salt-water plants take possession. Except for the burial and preservation of wood during the time of mangrove growth, the final stages may not be greatly different, so far as sediments are concerned, from other marine swamps (Vaughan).

Marine-swamp Deposits of the Past. The earliest marine-swamp deposits are those shown in some of the coals of the Mississippian and Pennsylvanian "Coal Measures," as evidenced by the association with marine fossils. There were swamps in the Devonian, but it is not certain that they were marine. It is suggested that many black shales of the geologic column had some connection with marine swamps.

FRESH-WATER SWAMPS

Fresh-water swamps are sequential to the inner parts of marine swamps (Fig. 11), but many fresh-water swamps, perhaps most, never contained salt water. They are connected with basins of varied origin and under some climatic conditions have nothing to do with basins and cover entire surfaces. These are usually flat or gently inclined, but they may be steep. Some fresh-water swamps contain large trees, as the cypress swamps of the South and the black spruce, cedar, and tamarack swamps of the North. Others support a low, tangled shrubbery, as the alder and sumach swamps.

Most fresh-water swamps of today have little chance of burial of their deposits. Such is true for the vast organic accumulations that cover large parts of the Laurentian Shield.

A swamp on a level or nearly level surface presupposes an initial high water table. A fallen and semidecayed vegetation serves as a blotter or sponge to raise the table. Such a swamp may raise its surface above surrounding levels and spread laterally. The shape would become a low dome, which under conditions of great rainfall may "burst" and spread the black organic matter over parts of the surroundings. These "bog bursts" are known to have produced streams of black soupy materials as much as 30 meters wide and more than 1 meter deep (Barkley).

Plants growing on flat-surface swamps consist of mosses, ferns, sedges, grasses, horsetail rushes, and various other plants. The tundra of subpolar regions is a flat-area swamp, with sphagnum moss the chief plant. Most swamps on flat surfaces have pools of more or less brown water scattered around in irregular distribution. The water in these pools may be extremely toxic to plants and animals.

Basin swamps are found about small lakes in which there is little wave activity, or about large bodies where shallow water or a barrier serves as a protection. The lakes may be of any origin, but the swamps tend to be better developed about lakes that are not of floodplain or delta origin, as these may receive so much mud as to destroy plant growth. Plants responsible for the accumulation of the organic materials are chara, sphagnum, canes, reeds, grasses, sedges, rushes, arrowleaf, and water lilies. Each type of plant is adapted to some depth of water, and they succeed each other from the shore outward. Some plants, as sphagnum, float on the surface; others, as the pond lilies and chara, are rooted in the bottom. Floating plants form mats that live above and die below. These catch sediments and in course of time may form a thick cover. Beneath the mat there is water or black sludge. The mat may move with wormlike motion when disturbed and is then termed a "quaking bog." In time it may support large plants and ultimately trees. These destroy the plants to which the mat originally was due. The basin in time becomes completely filled with semidecayed organic matter. Figure 12 shows a small lake invaded by plants.

As vegetation is advancing into a lake from the shores, the central part of the lake remains open and to some degree agitated. Bacterial action continues in these central waters, as agitation permits entrance of oxygen and prevents extreme toxicity, and bordering vegetation limits entrance of inorganic sediments. The organic deposits beneath this central part consist of waxes, fats, gums, and lignin, and there are frequently many tests of diatoms, producing an organic deposit quite different from that of toxic waters.

The above considerations do not apply to basins on deltas and river floodplains. These have different histories, as the occasional floods wash out toxic waters, make deposits of mud, and introduce river organisms. The organic sediments are rarely of the purity of those of other fresh-water swamps and layers of clay, silt, and, rarely, sand separate them. As these swamps very largely receive their water from streams, there is the possibility of frequent desiccation and mud cracking. Such does not seem likely in other fresh-water swamp deposits.

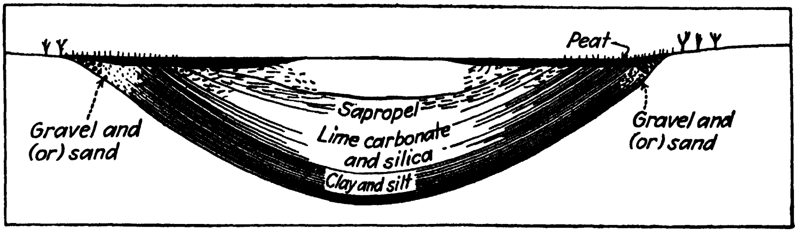

FIG. 12. A small lake invaded by plants. Water remains uncovered in the center and beneath a part of the accumulations of peat. The bottom is filled with deposits related to the character of the waters entering the lake and the plant life growing about the margins and in the lake.

SWAMP DEPOSITS

Swamp deposits in considerable, and in many cases in large part are of organic origin. Plant contributions are mainly grown in place and autochthonous. River swamps contain much clay, silt, and sand, all more or less black, owing to organic matter. Swamp deposits may contain iron oxide, which is rarely pure and usually contains a high percentage of phosphorus and considerable manganese oxide and silica. The beds of iron oxide are generally without stratification and may attain a maximum thickness of about 1 meter, but they are usually thinner. Many have concretionary structure. These are the bog-iron ores, known as *murram* in the tropics and *mocarrero* in Cuba. In some places they form the *iron pan*, a form of hardpan. The silica may be largely from diatoms. Thin beds of iron carbonate may also be present. Deposits of fresh-water swamps contain little or no iron sulphide. There may be considerable in marine-swamp deposits.

Some of the coal deposits are certainly of fresh-water swamp origin, and this can readily be proved for many of the coal beds of the Pennsylvanian system of the Appalachian Region, the Cretaceous of the Rocky Mountain Region, and the Tertiary of the Great Plains Region and the Puget Sound Depression.

The Lacustrine Environment

INTRODUCTION

Five varieties of lake basins—glacial, river, seashore, deflation, and crustal movement—seem to be most important.

Lake processes are physical, chemical, and organic. Physical processes are due to waves and currents, which in most lakes do not have the importance they do in the sea because of the absence of tides and the difficulty of producing large waves in small bodies of water. Waves and currents in most lakes have little or no effect a short distance below the surface.

Inorganic chemical processes do not seem to have much importance. Doubtless there are chemical and colloidal reactions leading to precipitation in lakes, and there are chemicophysical reactions connected with rise of temperature and agitation that cause escape of carbon dioxide and precipitation of carbonates held in solution.

Organic processes are important. On the basis of nutrient matter and oxygen in their waters, lakes may be placed in three classes: *eutrophic,* with little oxygen in the bottom waters and much nutrient matter; *oligotrophic,* with considerable oxygen in the bottom waters and limited nutrient matter; and *dystrophic,* with scanty oxygen in the bottom waters and little nutrient matter. The life is adapted to these different conditions. Shallow waters within the zone of agitation are inhabited by mollusks and crustacea. Plants thrive more or less abundantly in the zone of light, and the open waters support a more or less large population of fish, animal plankton, diatoms, and algae. Bottom waters beneath the zone of wave and current agitation may have a limited benthonic population of not a great many species. The plant life produces important deposits in the shallow waters and precipitates various carbonates in photosynthesis during life. Shelled organisms add to the carbonates, and their bodies add nutrients to the waters and organic deposits to the bottoms.

Certain organisms seem never to have lived in lakes. Among these are brachiopods, cephalopods, corals, echinoderms, and pteropods. Gastropods and pelecypods have probably dwelt there since the Middle Paleozoic. The fish of lakes are varied in kinds but can hardly be used to identify lake deposits, as some, like the salmon, in their life history go from fresh to salt water and return again to fresh water to die, and others, like the eel, go from salt to fresh and return. Some fish seem to become easily adapted, as sharks, for example, normally marine, live in Lake Nicaragua.

Most macroscopic benthonic organisms of lakes do not seem to live in depths greater than about 25 ft., and neither number nor variety is particularly large. Only small macroscopic organisms live on deeper bottoms, except that fish live on bottoms of lakes of temperate regions at the times of overturn and for some time thereafter. The plant life of lakes is abundant and ranges from bacteria, diatoms, and microscopic green algae to higher plants. Important among the higher plants are arrowleaf, water lilies, cattail, and pickerel weed, and sphagnum is a most important moss.

Lakes have no true littoral zone, as there is no tide, but changes of level connected with precipitation and wave action may produce conditions not unlike those of the ocean littoral. Shallow lakes with weak waves have little or no erosion of shores, and the shallow bottoms tend to become colonized by aquatic plants. Shores attacked by strong waves are eroded, and the materials of the bottom grade outward from gravels or sands at the shores to muds at little depth. Colloidal and fine sediments do not readily flocculate in lake waters; hence they are likely to be spread widely over lake bottoms. The deposition of fine materials tends to be essentially continuous over lake bottoms during those times of the year when there is no coat of ice, and the sediments are usually mixed with much organic matter and do not settle firmly on the bottom but exist as a sort of thick soup that is worked over by organisms so that stratification cannot develop. Stratification develops in the silts, sands, and gravels of the shallow water. The general absence of tides limits formation of currents, and this does not favor the formation of current ripple marks except as these are made by wave currents. Wave ripple marks may be common.

The foregoing are general characters. The following relate to particular lakes. Glacial lakes initially probably received glaciolacustrine sediments, and these pass into sediments that are reactions to the conditions following disappearance of the ice. Many glacial lake basins were shallow, and thus the typical lake deposits were shortly succeeded by deposits of swamps. Basins tend to be circular but may be elongated or irregular in outline.

Lakes connected with rivers are shallow and elongated. Many are built on channel deposits, and all are surrounded by floodplain deposits. The organisms dwelling in river lakes are derived largely from rivers, and they receive a new supply at each flood. River lakes frequently dry up or become filled with stagnant waters during intervals between floods, with consequent destruction of the whole or large parts of the animal population. In course of time the lakes are certain to be filled with sediments partly derived from floods and partly from organisms. The deposits are muds and sands, commonly well-laminated, usually black, and frequently highly mud-cracked. Organic remains may be common.

Seashore lakes are developed along the seashore. This is done through building of barriers by waves or, on deltas, by rivers. They contain either fresh or salt water and may begin as salt water and become fresh. They usually are shallow, and streams commonly flow into them. Materials are brought to these lakes by streams, by winds, and by waves over the barriers. There is also deposition of organic matter. The deposits thus present a mingling of the sediments of marine and fresh-water origin, and latterly and vertically may pass into the deposits of paralic swamps.

Deflation basins are produced in arid regions, and if they contain water it may be salty. They may contain organic materials. The sediments are clays, silts, sands, gravels, and evaporites. The fine sediments before lithification may be black because of formation of the monosulphide of iron.

Lake basins of crustal origin have a wide range of area, but none is apt to be small. Examples are the Great Lakes, Lake Baikal, and Lake Nicaragua. Waters may be fresh or salty. Shallow-water deposits of very large lakes are not quantitatively different from those of the sea, but those of deep waters consist of fine muds that seem to contain little or no molluscan life (Kindle, 1925). Salt-water lakes may make deposits over deep bottoms that are black from monosulphide of iron as seen on the bottoms of the Black Sea and Lake Baikal.

Fresh-water lakes may be classified on the basis of the composition of the contained water into hard- and soft-water lakes. Naturally, there are all gradations. Lakes containing soft water have a limited population of organisms requiring calcium carbonate for shells. Those containing hard water may have such organisms in considerable abundance. Deposits of calcium and magnesium carbonates are not formed in soft-water lakes, but bottoms to which muds and sands are not brought in considerable quantity may have deposits of nearly pure fine-grained organic matter which may be so lush in small lakes of woodland areas as to render bottom waters toxic. Hard-water lakes may receive deposits of calcium carbonate over bottoms to which little mud or sand is brought, and the calcium carbonate may accumulate to depths of water exceeding 60 meters (200 ft.) as is the case in Green Lake in east central Wisconsin (Twenhofel, McKelvey, Carter, and Nelson; Twenhofel and Feray).

STRATIFICATION OF LAKE WATERS

The density of water varies with the quantity of material in solution and the temperature (Kindle, 1927). For distilled water the temperature of greatest density is 4°C. and at some other figure if the water has material in solution. If fresh water is cooled at the surface, the chilled water sinks and forces warmer and lighter water to the top to be chilled in turn, and this continues until the water of a lake has attained throughout the temperature of greatest density. Further cooling may result in freezing of surface waters, since with lowering of the temperature below 4°C. the waters become lighter and no longer sink. Meanwhile there has been a complete *overturn*, and waters provided with oxygen have gone to the bottom from the surface, and bottom waters depleted of oxygen and filled with carbon dioxide have been brought to the top. After an overturn the waters of a lake are provided with oxygen throughout. Spring melting or the ice again raises the surface temperature to the position of maximum density, and again there is an overturn.

Seasonal overturn does not take place in lakes of tropical regions, but irregular overturn may take place following a period of cool weather. Lakes of subtropical regions may have an overturn during the winter season, but as the weather in such climates, illustrated by Florida, has short periods of cool and even relatively cold weather, the waters may overturn several times during a winter. Lake waters are said to have *direct thermal stratification* when the colder and heavier waters are on the bottom. Bottom waters of tropical deep lakes tend to be foul from lack of oxygen and presence of products of decay. There is an abundance of carbon dioxide which favors solution of any carbonates entering the bottom waters and makes accumulation difficult if not impossible. Bottom waters of lakes of polar regions are at the temperature of greatest density, but the waters at the top are colder than those below. There is no regular overturn, and the thermal stratification is said to be *indirect.*

Thus, fresh-water lakes may be placed in three general classes:

1. Lakes of tropical regions with no regular overturn, colder waters on the bottom, warmer on the top, foul waters on the bottom, direct stratification.
2. Lakes of polar regions with no overturn, warmer waters at bottom (but cold), coldest on top, indirect stratification.
3. Lakes of temperate regions, with autumn and spring overturn, direct stratification in summer and indirect in winter; bottom waters carry oxygen after overturns, are foul after oxygen exhausted.

Upper waters of lakes are subject to greater or less daily changes of temperature, but they have a more constant temperature than the overlying air, owing to mixing and stirring by wind and convection currents. These upper waters of more or less variable temperature vary in thickness with the conditions and form a zone known as the *epilimnion* (Birge and Juday; Kindle, 1927). The bottom waters of a lake are protected against daily temperature changes and maintain a relatively constant temperature in tropical and polar lakes and in the long intervals between overturns of temperate lakes. These lower waters have lower content of oxygen, higher content of carbon dioxide, and higher hydrogen-ion concentration (low pH) than the epilimnion. This zone is known as the *hypolimnion.* Between the epilimnion and

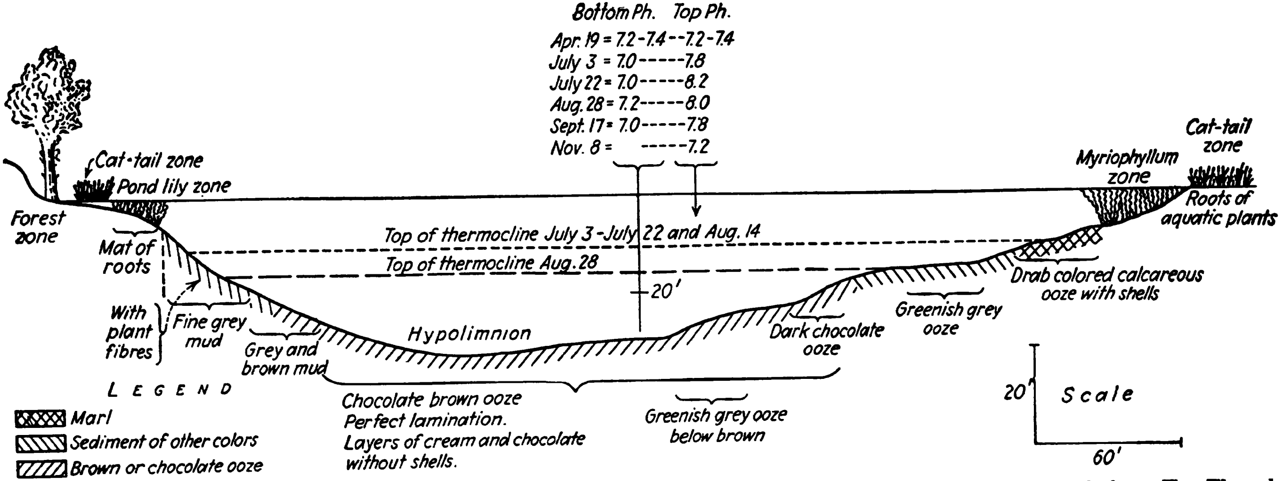

FIG. 13. Section across McKay Lake showing distribution of bottom sediments. The cold hypolimnion has relatively low pH. The calcareous drab-colored ooze is deposited in the warm alkaline waters. *Elodea* as well as *Myriophyllum* thrive in the *Myriophyllum* zone. (*After E. M. Kindle, Trans. Roy. Soc. Canada*, 21 (1927), 5.)

hypolimnion is an intermediate zone, termed the *thermocline*, which is arbitrarily defined as that zone in the water of lakes in which the change of temperature exceeds 1°C. per meter of depth. The thermocline on the basis of temperature is somewhat sharply defined from the epilimnion and passes gradually into the hypolimnion. Much organic matter passes through the hypolimnion in transit from the upper waters to the bottom. This consumes oxygen as it decomposes and produces carbon dioxide. As organic matter is not abundant in polar lakes, the consumption of oxygen in the hypolimnion is not great, and hence the waters of these lakes do not commonly become so foul as the bottom waters of temperate and tropical lakes (Fig. 13).

The conditions in the epilimnion of temperate and tropical lakes are favorable for the maximum development of life and for oxidation. There is an abundance of oxygen for animals and adequate carbon dioxide for plants. The hypolimnion of small lakes commonly has colder waters, higher carbon dioxide content, lower content of oxygen, and lower pH than the epilimnion. These conditions favor reduction of inorganic matter, solution of carbonates, and scarcity of life. The waters of the hypolimnion of tropical lakes may be so acid that permanent deposition of carbonates may not be possible.

TRANSPORTATION IN LAKES

Transportation in lakes naturally varies with conditions. There are currents due to winds, to inflow and outflow of streams, and to differences in densities of water from temperature and material in solution or suspension. Muddy waters flowing into lakes, particularly those of warm or temperate regions, tend to sink to the bottom where the waters mingle with the colder waters of the hypolimnion, and if there are no waves to keep the sediments in suspension, deposition takes place at the contact of the muddy waters with the clearer and quieter waters of the hypolimnion. This deposition may form ridges on the bottom. This takes place where the muddy water of the Rhine flows into Lake Constance, or similar water of the Rhone enters Lake Geneva. Muddy water after sinking to the bottom may become a density current to the deepest part of a lake basin and carry its sediments to the deepest bottom. Water with a large content of dissolved matter may also become a density current if it enters a lake whose waters have little matter in solution. Warm waters flowing into lakes containing cold water float on the surface, and contained sediments become generally distributed throughout the epilimnion and ultimately are spread somewhat uniformly over the entire bottom of a lake. Fresh waters flowing into salt waters also float.

Lakes that are fed by melt waters from glaciers present special cases of transportation. The melt waters are at, or just above, the freezing temperature and thus on the low-temperature side of the condition of greatest density. Frequently a glacier is partly submerged in the waters of a lake. If the melt waters enter the lake at the surface, they float; if below the surface, they rise to the surface. Strong anticyclonic winds blow from glaciers, and these direct the melt waters away from the ice and spread them widely over the epilimnion, so that the burden of sediments is dropped quite uniformly over the bottom. The circulation of the waters in a glacial lake is shown in Fig. 14.

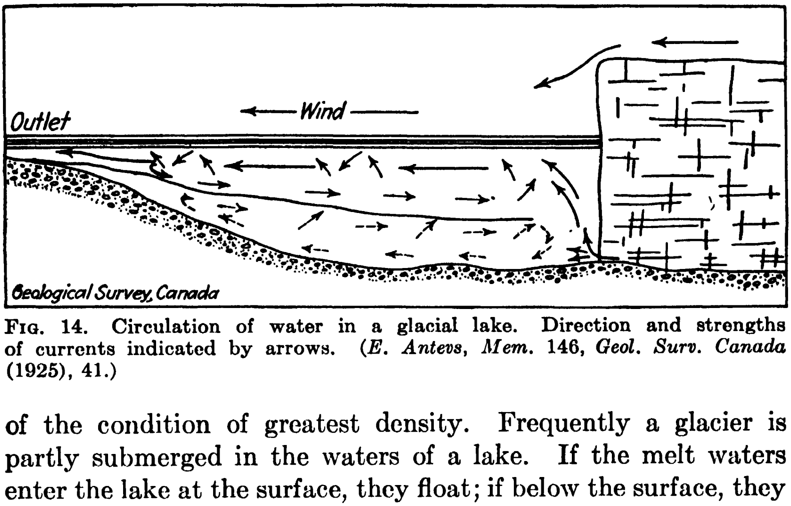

FIG. 14. Circulation of water in a glacial lake. Direction and strengths of currents indicated by arrows. (*E. Antevs, Mem.* 146, *Geol. Surv. Canada* (1925), 41.)

THE LAKE CYCLE

Lakes are transitory features. No sooner do they come into existence than they begin to disappear. A basin is filled through deposition of sediments derived from streams, erosion of shores, etc., and an outlet may lower the channel through which it flows, so that the bottom rises through deposition and lake level falls through erosion. The shore thus migrates centralward, spreading shore deposits over those previously made as it retreats. In time a lake either passes into a swamp or becomes filled to the level of the outlet with inorganic sediments. The outlet stream then extends itself headward into the lake deposits and ultimately connects with streams draining into the lake. If the level of the base level of erosion lies below the original bottom of the lake, all the lake sediments are ultimately removed, and the lake cycle is complete.

LAKE SEDIMENTS

Lake sediments consist of marls, tufas, clays, silts, sands, gravels, iron hydroxides, iron carbonate, silicon dioxide, manganese oxide, calcium phosphate, organic matter, and evaporites.

Lake marl is a mixture of calcium carbonate with a variable content of impurities. Colors are various. Most marls are gray to white or pale blue, but red and black marls are not uncommon. Some marls have the calcium carbonate in the form of broken shells, whereas in others there is no evidence of shells. Marl has been stated to form only in the epilimnion, and this may be the case in some lakes, but bacteria that precipitate calcium carbonate may be present in muds beneath the hypolimnion to depths of water of 25 meters, and it is known that extensive deposits of marls have been made to greater depths (Fig. 15). Marl of which a large part is composed of shells of gastropods has accumulated over the bottom of Green Lake in eastern Wisconsin to depths greater than 60 meters (200 ft.). Extensive deposits of marl have been made in many existing and extinct lakes of Wisconsin, Minnesota, and other states in northern United States and in Canada where the underlying rocks are limestones. Marl deposits are not made in soft-water lakes, and it does not seem to be a common sediment of large lakes (Davis, Thiel).

Tufa is a lime carbonate deposit that is more or less porous and banded. It seems to be quite largely due to algae, but evaporation may play a part. The most extensive tufa deposits in the United States are those in the remnant lakes of former Lake Lahontan in Nevada (Russell).

Lake clays, silts, sands, and gravels are extensive in some lakes. Sands and gravels are largely confined to shallow depths. Clays and silts may be spread over any lake bottom to any depth. The clays and silts range in color from white to gray-blue and gray-green to black. Sands are generally gray, but

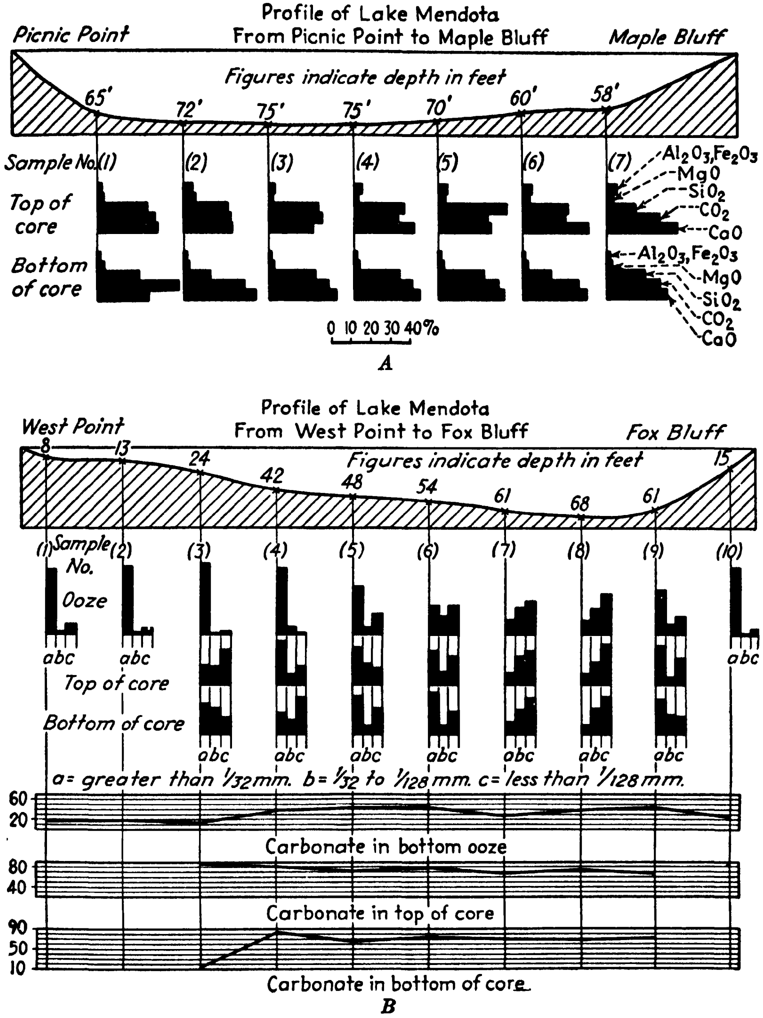

Fig. 15. Profile showing composition of the sediments on and beneath the bottom of Lake Mendota, a hard-water lake adjacent to the campus of the University of Wisconsin, Madison, Wisconsin. The length of the profile is about 2 miles. The silica shown in 15*A* is partly due to the tests of diatoms. 15*B* shows the grain size of the particles and the variation in carbonate content with depth beneath the bottom. The core is thought to represent a thickness of 4 to 5 ft. The $CaCO_3$ is soft marl with few shells.

If a similar profile and group of analyses were made of a soft-water lake in the woodlands of northern Wisconsin the sediments would contain little or no calcium carbonate and would contain a high content of organic matter.

all clastic sediments may be dark from incorporation of organic matter. Iron hydroxides are common sediments in some lakes, as certain lakes of Finland, Sweden, Russia, Canada, Maine, Wisconsin, etc. Limonite has been deposited to a thickness of about ½ meter in depths of about 10 meters in Swedish lakes. The deposits are more or less nodular and concretionary and are of irregular distribution and are said to be more abundant where there is lush growth of vegetable matter. The colors are some shades of black and brown, and some vegetable matter and gelatinous silica are usually present. They also contain considerable phosphorus in the form of earthy vivianite, and there is usually some manganese oxide. Deposits of iron and manganese oxides in Trout Lake in northern Wisconsin form a crust at least 15 cm. thick in depths of water to 26 meters (86 ft.). Analyses show that the deposits contain an average of 16 per cent metallic iron and 30 per cent metallic manganese. There are also considerable organic matter and some silica. Macroscopic plant life is not known to be present at the places of deposition of the iron and manganese oxides (Twenhofel, McKelvey, Nelson, and Feray). Iron carbonate is present in the deposits of some lakes as nodules and in disseminated form. The quantity usually is small and seems to be related to richness in organic matter. While manganese oxide is generally associated with iron oxide, it may also occur alone in relatively pure state as nodules and crusts, and in Ship Harbor and other lakes of Nova Scotia it is in the form of saucer-shaped disks more than 5 cm. in diameter (Kindle, 1931) (Fig. 16). Iron sulphide is a not uncommon constituent of the deposits of salt-water lakes and may be present in the deposits of some fresh-water lakes. Peat is a common sediment about the margins of many lakes, and central parts of lake basins not receiving much clastic sediment may become filled with black muds rich in organic matter (Rickett). Hard-water lakes do not commonly contain such deposits in very pure form, but small soft-water lakes may contain a great deal. Woodland lakes of northern Wisconsin (Twenhofel, McKelvey, Carter, and Nelson) contain large volumes of organic sediments of plant origin. These are deposited beyond the shallow bottoms of the shore zone and consist largely of a pale greenish-yellow gel in which are fragments of plants, diatoms, sponge spicules, and inorganic matter, the proportions varying considerably from place to place. Deposits of this character are nearly 16 meters (52 ft.) thick in Allequash Lake in northern Wisconsin. The water content is about 90 per cent, and there is little variation no matter what the depth. After drying, the organic sediments shrink to a very small percentage of the original volume and become black, tough, and rubbery. The dried sediment ignites in an explosive manner. The 16 meters of thickness of these organic sediments in Allequash Lake would make not more than the 1.5 meters of solid sediments. Such fine organic sediments of lakes have been termed *sapropel* and have been differentiated into *gyttja*, in which the organic matter is more or less determinable, and *dy* if the organic matter before precipitation is in colloidal form (Naumann). Gyttja is said to be characteristic of eutrophic and oligotrophic lakes, dy of dystrophic lakes. Sapropel may contain considerable calcium carbonate. Gelatinous silica is doubtless an original precipitate in lake deposits from reactions of colloidal silica with electrolytes. Tests of diatoms are common, and diatomite is a not uncommon lake sediment. Evaporites are characteristic deposits of lakes of arid regions. Figure 13 shows the distribution of sediments in McKay Lake near Ottawa, Canada.

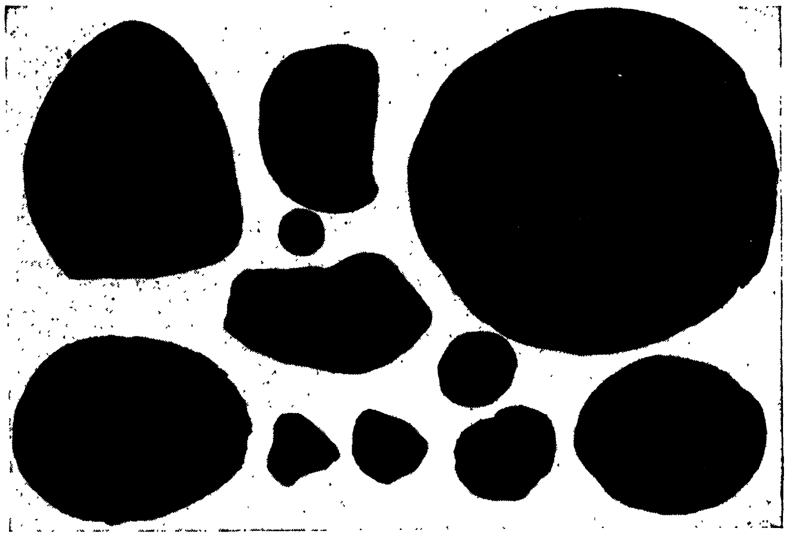

FIG. 16. Disk-shaped concretions of manganese oxide from the bottom of Ship Harbor Lake, Nova Scotia. (*Photograph from E. M. Kindle, Geol. Surv. Canada.*)

Lake deposits are little known in the geologic column. It is possible that parts of the Pre-Cambrian are of lake deposition and perhaps some parts of the Old Red Sandstone. Many lake deposits of small basins are probably present in the "Coal Measures" of the Pennsylvanian, Jurassic, Cretaceous, and Tertiary systems. The Green River shales of Wyoming and Colorado seem to be lake deposits, and the various Tertiary basin deposits of western United States were in some cases deposited in lakes, as was the case in old Lake Florissant in Colorado.

The Cave (Spelean) Environment

Caves may be large when considered as cavities in the ground, but their magnitude compared to other sedimentary environments is small, and deposits made in them are relatively insignificant. The importance of cave deposits is qualitative.

Caves probably have been places of refuge for animals ever since there was land animal life, and today caves constitute the most fertile place to find remains of man and contemporaries, as some of the latter took refuge in the caves with him and others were brought in for food. Other animals before the advent of man probably had like relationships. Some animals, such as bats, spend the days in caves, and a few make permanent homes therein. Some of the latter are eyeless or have nonfunctional eyes.

Caves may be developed by both water and wind. Caves eroded by water carry evidence of the agent of erosion on the walls and roof in the form of solution and decompositional effects and deposits made by waters seeping into them. Wind-worn caves are due entirely to physical erosion and are open widely to the outside; the walls and roof do not carry decompositional and solution effects.

Deposits in caves are made from rocks falling from the roof and walls, by waters that enter unsubmerged caves in any way, and by the waters of a submerging body of water. The deposits are of both physical and chemical origin. Physical deposits are represented by debris that has fallen from roof and walls, which may pile up to depths of several feet; by small deposits of clays, silts, sands, and gravels introduced by underground streams; and by clays, silts, sands, and gravels brought by submerging waters. Roof falls and stream action can hardly fill a cave, but submerging waters may readily do so, and thus there is produced the anomaly of a younger marine deposit laid down beneath an older one.

Chemical deposits in caves are the various stalactites, stalagmites, and similar substances. These are commonly composed of calcite, which is banded and forms rocks variously known as onyx and travertine. Rates of growth of some stalactites have been determined, but there are so many variables controlling growth that figures have little general application. Determined rates of growth range from a few centimeters per annum to a little more than 48 cm. The last is phenomenal (Allison). Stalactites and stalagmites may join to form pillars, and in course of time a cave may become filled with these deposits to make a solid deposit of onyx. It is possible that a cave may be submerged and become filled with marine chemical deposits, but proof of the existence of such has not been forthcoming in spite of the efforts of some geologists to explain certain deposits on this basis.

Caves occasionally contain organic deposits consisting of the bones and excrements of inhabiting animals and their prey. Cave guano made from the bones and excrements of bats is perhaps the greatest in quantitative importance.

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TRANSITIONAL ENVIRONMENTS

The place of meeting of land and sea produces environments that are ruled jointly by the two great environmental realms, or each alternately maintains an often repeated rule for a short time. These transitional environments thus have sediments that form either from the combined impact of the land and sea, or sediments dominantly due either to one or the other of the great environmental realms. There are five such transitional environments of which one, the paralic swamp environment, has already been considered. The other four are the littoral, delta, estuary, and marginal lagoon. The littoral is the arena of contest of land and sea, with the land generally retreating and marine processes dominant. In the delta, on the other hand, the sea is retreating during the early stages of a delta cycle and advancing during the later stages of the cycle. The result may be intimate dovetailing of marine and continental sediments. The estuary is a place of divided rule, a part belonging to the sea at high tide and to the land at low. The marginal lagoon is a place of more or less united rule with the waters ranging from fresh to salt. The sediments deposited in the littoral, the marginal lagoon, and the estuary come both from the land and the sea; those on the delta come largely from the land.

THE LITTORAL ENVIRONMENT (BARRELL, 1906; MARTENS)

The littoral environment consists of that part of the shore zone that is exposed at low tide and covered with water at high tide. The upper margin is placed at the highest limits of waves. The lower margin is indefinite.

Ordinarily the area of littoral environment is only a few meters, or at most a few tens of meters, wide. There are places of greater widths, as in the Bay of Fundy, and particularly in Hudson Bay where the tidal zone is 6 to 8 miles wide in places and a boat landing at low tide may not see the shore of high tide. The North Sea has a wide tidal flat in the Wattenmeer on the north coast of Germany (Hantzschel). Some of the epicontinental seas of the Paleozoic may have had conditions that made for high tides and a wide littoral zone. There certainly were extensive lowlands over which high tides could have advanced for long distances. The present area of the littoral environment is placed at 62,500 square miles. It may have had greater area during some geologic periods.

The littoral environment has the unique characteristic of exposure of the bottom for a longer or shorter period of time twice each day, and when it is submerged it is always within the influence of strong wave action. The environment is a rather difficult one in which to live. Organisms must either be firmly attached to the bottom, as the black mussel, *Mytilus*, and the acorn barnacle, *Balanus;* burrow in the mud, as the littleneck clam, *Mya;* or live in holes or underneath large rocks. Some animals, such as the sea urchin, dig holes in solid rock. Each of these organisms must also have structures and arrangements for keeping the body wet, either by closing the shells, retaining water in the holes, or burrowing in the muds and sands. Those that burrow take great risks, as high evaporation may increase the salinity of the waters in the muds and sands, a heavy rain may dilute them, and a hot day may heat them to lethal temperatures.

The more or less common wave action abrades and grinds the rocks and shells present, and these are reduced to rock flour, "sands," and larger particles.

The materials forming the surface of the littoral environment vary from place to place. In some places the surface is bare, more or less polished rock. In other places it is covered with boulders, gravels, sands, muds, or entire or broken shells. In places where waves are not provided with rocks with which to abrade and wave action is not too intense, the surface may be covered with dense mats of marine plants. Mangrove shores are one phase of the littoral environment and coral-reef shores another.

The dominant processes in the littoral environment are physical; chemical and organic processes are subordinate. Sediments are acquired from the shore and shallow water and transported and deposited almost entirely by physical agencies. Organisms precipitate materials from solution, but unless the substances thus formed are almost immediately cemented to the bottom, the waves and the currents acquire them and become the agents of deposition. The little chemical action consists of the effects of water upon materials of the bottom and shores and of evaporation between tides. The former is rarely apparent because of the intense physical action, and the products of evaporation in most cases are returned to solution on return of the tide. Organic processes obtain where conditions are favorable, as shown by the local abundance of such shelled animals as the black mussel, *Mytilus*, the acorn barnacle, *Balanus*, and dense mats of seaweeds.

Littoral deposits consists largely of materials in the range from boulders to clays. Locally there are shells, which in most cases are broken, so that these are perhaps better termed shell "gravels," "sands," or "muds." Locally, seaweeds live in abundance in the littoral, and at times waves bring tons from shallow waters to the shores. Few of these are likely to make permanent deposits. In places a single kind of deposit, such as gravels, sands, silts, or clays, may cover the littoral for miles, and elsewhere the materials may be jumbled together in extremely irregular arrangement with lenses of clay, silt, or sand in the midst of gravels or boulders, so that within a distance of a few tens of feet one may pass over sediments in the range from boulders to muds. In general, the coarsest materials and materials of highest specific gravity are on the upper part of a beach, and if beach materials should be composed of magnetite, garnet, and quartz sands, the magnetite sands would be on the highest part of the beach and be followed in order downward by garnet and quartz. Particles in the range from sands to boulders may be well rounded, but there are many accumulations of boulders, gravels, and sands in the littoral of which the particles are little rounded and even angular. Some shores margining lowlands are composed entirely of carbonaceous materials, and this type of deposit is common on low coasts protected by very shallow water and affected by weak tides. It is suspected that some of the widespread black shale formations like the Chattanooga and associated shales of Kentucky, Tennessee, and Ohio developed in the littoral and adjacent shallow waters of either a retiring or an invading shallow sea. If there is enough wave action to keep shores clean of vegetable matter and lands are low, so that clastic sediments can not enter, and waters are congenial, calcareous sediments accumulate.

Colors of littoral sediments are in the range from light to black. Some black sediments, as the black sands on the coast of the Wash of England (Kindle), are due to coatings of the black monosulphide of iron. Others may be due to manganese oxide. Clastics tend to have light colors unless composed of black minerals or rocks.

Mud cracking is not uncommon over high parts of the littoral environment in sediments that permit formation. Rain impressions may be made on all muds and sands; bubble impressions in muds; and sand holes are made and as rapidly destroyed as the waves advance to and recede from shores. Both wave and current ripple mark may be present on all sands, the latter the more common, and the rhomboidal ripple mark that resembles the overlapping scales of a ganoid fish may be extremely abundant. Rill marks are extremely common, and cross-lamination may develop in all sands. As winds frequently drift sands of beaches, wind cross-lamination may be present. However, each high tide may rework the beach materials, obliterate all previous markings, and remake many of them anew. Sorting ranges from extremely excellent to very poor.

Both land and marine animals come to the littoral for food, and crows and gulls follow retreating tides. These make tracks on all sediments where that can be done. Annelids and other organisms burrow through the muds and sands and form burrows that range in size from very small to 1 cm. or more in diameter. Some are nearly vertical tubes, as in the case of *Mya*, and U-shaped tubes are made by *Arenicola*. Some animals make horizontal burrows. Gastropods and crabs leave trails of various kinds, and at low tide the tidal zone is frequently crisscrossed by large numbers of trails.

The thickness of littoral deposits for a stationary sea level can be little greater than the tidal range at each place. This is usually less than 5 meters. If sea level rises, the littoral environment moves inland and acquires a higher position with respect to the original position, but the littoral deposit becomes no thicker unless the conditions are such as to make it possible for the new littoral deposit to be built on the old. This is not likely. Furthermore, it is probable that the coarse materials of the old littoral deposit are moved to become a part of the new deposit, and the fine materials are moved offshore to deposition on neritic bottoms.

As a part of the geologic column, littoral deposits have small quantitative importance. Only under exceptional conditions can they have a significant thickness. They tend to be destroyed with each tide and are transported inland to make new littoral deposits on rise of sea level. A falling sea level leaves a littoral veneer over the surface abandoned, but this is never thick, and weathering and erosion may soon destroy it. The importance of littoral deposits lies in the fact that they record the former position of a shore line. They have often been identified, but it is probable that many of these are shallow neritic deposits and that littoral deposits are rare. Deposits of either the littoral or upper part of the neritic environment should be present at the base of every marine formation that is separated from the strata below by an unconformity due to retirement of the sea.

THE DELTA ENVIRONMENT (FOX, JOHNSTON, RUSSELL AND RUSSELL, SYKES)

General Considerations

A delta is defined as the deposit of sediment made by a stream at the place of its entrance into a permanent body of water. Waves and currents of the permanent body of water aid in deposition. The sediments are supplied by the stream. A delta increases in dimension when a stream supplies more materials than the waves and currents of the permanent body of water can carry away, and it tends to decrease in extent when the reverse obtains. The permanent body of water may be a lake or the sea; the former is not considered here.

A delta is composed of sediments of subaerial and subaqueous deposition. The former are fluvial, lacustrine, paludal, and in some cases aeolian. The subaqueous deposits are mainly of fluvial deposition, but some are deposited by waves and currents of the body of water into which the stream empties. The subaerial deposits are made on that part of the delta that is exposed most of the time and is under water only during floods.

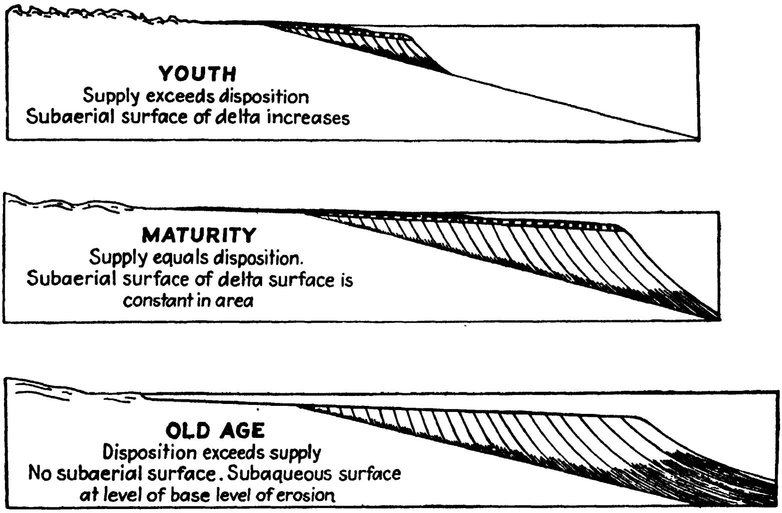
YOUTH
Supply exceeds disposition
Subaerial surface of delta increases
MATURITY
Supply equals disposition.
Subaerial surface of delta surface is constant in area
OLD AGE
Disposition exceeds supply
No subaerial surface. Subaqueous surface at level of base level of erosion

FIG. 17. Diagram showing components of deltas. The black indicates subaerial topset beds, the gently inclined with spheres the subaqueous topset beds, the distantly spaced the forest beds, and the closely spaced the bottomset beds. (*After J. Barrell, Bull. Geol. Soc. Am.*, 23 (1912), 396.)

From a physiographic point of view a delta has three parts. There is an upper part that is quite flat and has the two parts of the *subaerial* and *subaqueous plains*. These are separated by a somewhat indefinite boundary known as the *shore face*. The sediments covering these two plains lie in nearly horizontal position and are termed *topset beds*. Extending outward and downward from the subaqueous plain is the *foreset* slope, upon which sediments, termed *foreset beds*, are deposited in positions paralleling the inclinations of the surface. These inclinations may be high and sharply defined from the topset beds in small bodies of water with weak waves and currents, but are low in large bodies of water with strong waves and currents, so that they pass by almost inappreciable change in inclination into the topset beds. Beyond the delta front the surface is of gentle inclination and is underlain by the sediments termed *bottomset beds*. The three components of deltas may readily be differentiated in small deltas made in laboratories and in those of small lakes made by melt waters of glaciers. There is great difficulty, if not impossibility, of recognizing the three components in large deltas, as these are really composed of small deltas at the mouths of distributaries or different places of entrance into the sea of currents. It seems likely that the three components may be identified in each of these. Natural levees border each distributary, and between distributaries there are areas on which water may be standing beneath which fine muds may be deposited. As the muds of deltas contain much water at times of deposition, settling will take place as the water is expelled. Russell and Russell state that on the Mississippi Delta the settling amounts to 8 ft. or more per century. Subsidence is certainly an important feature of all delta deposits as most muddy sediments may contain water to an amount exceeding 50 per cent.

Lagoons may border the subaerial plain on the seaward margin, and deposits made in these have the characteristics of that environment except that the sediments may not pass upward into those of swamps, but into muds and sands brought by streams.

Sediments composing the topset beds of the subaerial plain consist of organic matter, clays, silts, and sands. Gravel may be present but does not seem to be a common sediment in deltas except in those of streams that flow into a sea or lake directly from uplands under which conditions gravel may become a considerable part of the sediments. Deposition over deltas takes place in channels of the distributaries, on the natural levees, and in marshes, lakes, and bays between distributaries. The distributaries extend outward and seaward or lakeward like the veins in a leaf (Fig. 18) and, as a delta is built upward, the deposits of each distributary are irregularly superposed on its preceding deposits, and thus the channel deposits rise more or less irregularly in a vertical section. Any vertical section through the deposits over a delta may pass through the channel deposits of one or more distributaries. The coarsest sediments deposited over a delta are likely to be in the channels. These may be cross-laminated, exhibit considerable slumping phenomena and much cut-and-fill structure. Because of the migrations of channels, floods, cut-and-fill, and irregular movements of stream waters, bedding is likely to be locally very irregular and variable. This has been described to some extent by McKee in the deposits of the Colorado River Delta. Mud cracking and rain and bubble impressions may form in the deposits of the subaerial plain where composition of the sediments and environmental condition permit.

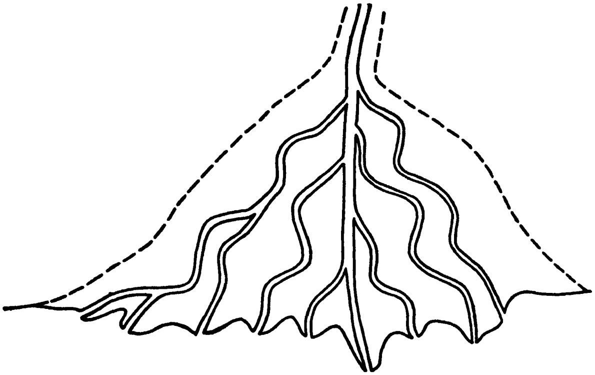

FIG. 18. Plan of the venation of a *Hibiscus* leaf from which the terminal end had been clipped and the venation changed from straight to crooked. The results may be considered to represent the subaerial portion of a delta built by a river and its distributaries. Suppose the entire system to be alternately shifted to the right and left in the manner of a pendulum but with great variation in the swings in either direction while at the same time each vein moved independently of the others. The progressive path of each vein would be like that of a distributary in a building delta which at the same time is settling under the accumulating load.

Colors of sediments in the subaerial parts of a delta range from gray to black, but red and yellow may be present under some conditions. Sorting is usually not good except on wave-washed beaches and in dunes. Remains of land and fresh-water animals may be abundant, and plants in place may be common. Remains of marine organisms are not precluded but should not generally be common. These owe introduction to birds, some land animals, and storm waves, or they come from marine animals that dwelt in tidal channels which often in tortuous courses cross subaerial and subaqueous plains.

The shore face separating the subaerial from the subaqueous plain may be well defined, but in many deltas the swamp vegetation of the subaerial plain may pass into that of the subaqueous plain without break of any kind. The detailed characters of the shore face and its presence or absence depend on the vigor of the waves, the extent of deposition, and the character of vegetation. Strong and vigorous wave action favors a steep shore face, which may extend for 1 meter or more below sea level. This feature is important, as it separates the deposits of standing water from those of streams.

The subaqueous plain receives sediments deposited by streams, waves, and currents. The waters range in salinity from fresh to salt, and the salt content varies from place to place and from time to time as the quantity and places of inflow of fresh water and the direction and intensity of tides, waves, and currents change. Sedimentary units tend to be discontinuous and to dovetail in a more or less intricate way. Some of the sediments may come from the sea, but most are derived from the land and are in the range from sands to clays and vary in sorting from fair to very poor. The coarsest sediments and also the best sorted are over areas of much wave wash. Inclinations of units trend outward in radial manner from the mouths of the distributaries. Marine organic remains may be present, particularly in protected areas between places of distributary discharge. Remains of land and fresh-water animals may also be present. Chemical sediments, among which are carbonates and glauconite, are possible on the outer borders. Cross-lamination may be common in sands. Plants should not be in positions of growth, but root structures may lodge in the positions in which they grew and thus simulate natural positions.

Foreset beds owe their deposition to stream and marine agencies with the former dominating adjacent to areas of distributary currents and the latter over intermediate areas. Angles of initial inclination are various, and in places of weak wave and current action they tend to be high, but in large bodies of water in which waves and currents have large spreading ability the inclinations may not be greatly different from those present in the sediments of the subaqueous plain. The upper part of the foreset slope has the sediments deposited to the level of the temporary base level of deposition; those on the lower slope are below that level. Those on the upper part may be subjected to much wave and current wash. This tends to remove the finest sediments. The sediments that remain are better sorted and somewhat coarser in grain. In general, the sediments of the foreset slope may have coarser grain than those of the subaqueous and the subaerial plains away from distributary channels.

Sediments of the foreset slope are composed very largely of sands, silts, and clays in lenticular units, with the lenticularity much more obvious in waters with weak waves and currents than those in which waves and currents are strong. There may be considerable local accumulations of calcareous sediments, and there may also be much precipitation of colloids of iron oxide and silica where fresh and salt waters mingle. The colloids may be disseminated in the clastic sediments or concentrated in concretionary structures. If deltas are not receiving much clastic material, the colloids may be precipitated in the form of beds and lenses. Certain chemical sediments, such as glauconite, may also form.

Colors of sediments of the foreset slope depend upon conditions, and shades of gray or blue should ultimately dominate, although, as in the deposits made in the sea by the Amazon River, the colors may be red. If organic matter is buried with sediments, as would be the case under most conditions, the red would be lost through reduction of iron.

Shell matter should be more or less abundant over foreset bottoms, particularly between distributary currents, and remains of fresh-water and land animals and plants from time to time may be buried with other sediments.

The areas of both the subaqueous plain and the foreset slope are environmentally unstable, as the currents of the distributaries shift from place to place and as stream discharge is variable. Shift of a distributary introduces fresh water where previously the water had been salt. Colonies of marine organisms may thus be entirely killed in a very short time and at the same time be buried in the positions in which they lived, or they may be killed merely by burial, as muds and sands may suddenly be deposited where for months the bottom may have received little sediment. This has been described for the deltas of the Fraser River in

British Columbia (Johnston, 1921), the Mississippi River, and some rivers of India.

The bottomset sediments are quite largely below base level of deposition and hence may not undergo much washing after deposition. They are also deposited under conditions that are largely marine. Terrigenous sediments consist mostly of fine suspended and colloidal materials of which large parts owe precipitation to flocculation by electrolytes and colloids of opposite sign. It is probable that much colloidal silica is deposited with these fine sediments. Sediments are believed to be laid down much more slowly than are those of any other part of a delta. The environment is quite stable, particularly if the bottom is deep, as the fresh waters flow outward over underlying salt water.

As a delta advances, bottomset beds are covered by foreset beds, and these in turn by topset beds, so that a vertical section through a delta that has been built under conditions of a stationary sea level first encounters the topset beds of the subaerial plain and in succession penetrates those of the subaqueous plain, the foreset slope, the bottomset beds, and typical marine deposits at the bottom.

It should not be assumed that the sediments of the different components are to any degree sharply defined either in attitudes of deposition or in degree of sorting or grain size. Sediments of equal coarseness may be found in topset and foreset beds. All contain sands, silts, and clays, except that sands are not common in bottomset sediments. Gravels are uncommon in any part of a large delta but may be present if there is a near-by source.

The delta of the Nile is thought to illustrate fairly well the physiographic features of large deltas (Barrell, 1912). The subaerial plain contains many lakes and swamps, lagoons border its seaward margin, and tortuous distributaries wander through it. The shore face is said to be distinct and to extend to a depth of 6 to 10 meters. The subaqueous plain extends with gentle slope outward to a depth of 50 meters, which is attained 50 km. offshore. There is then a more rapid descent to 200 meters and a still more rapid descent to 1,000 meters. The bottom from 50 to 200 meters is placed in a transition zone between the subaqueous plain and the foreset slope. The bottomset beds extend outward and downward from the depth of 1,000 meters.

Extensive studies have been made by Russell and Russell of the physiography, sedimentary conditions, and sediments of the subaerial part of the Mississippi Delta. Natural levees of distributaries, which rise about 4.5 meters (15 ft.) above sea level in the vicinity of New Orleans, 3.5 meters (12 ft.) in the neighborhood of Point a la Hache (about halfway to the passes), and about 1.5 meters (5 ft.) at Head of Passes extend seaward as subaqueous levees. Wide expanse of marsh lies between the subaerial natural levees. These are at about the level of ordinary high tide and are penetrated by tidal streams which connect ponds and lakes of the marshes, carry salt or brackish water into these, and by intricate network lead to coastal indentations. Subsidence takes place due to compaction of sediments, with major settling occurring at places of major deposition. The subsidence creates delta flank depressions which become invaded by sea water and extend marine sedimentation inland from areas of contemporaneous continental sedimentation. Subsidence in Bernard Parish is supposed to have decreased the land area by 63 square miles. Deposition by the river has added 77 square miles to the land area of Plaquemines Parish, but the net result of deposition and settling during the past century seems to show that increase and loss of land area just about balance.

The sediments of the lower delta of the Mississippi River show marked contrast between those of channel bed, natural levee, marsh, lake, and bay and, as these surface features are subject to frequent change of position, it results that a vertical section through the delta is almost certain to pass through several changes of the various kinds of sediment. The detrital constituents of the delta sediments have an upper size limit of about 1 mm., and the range is thence downward to colloidal dimensions. Analysis of samples from over 300 surface deposits and bore holes gave the following:

Coarse sand	$\frac{1}{2}$ to 1 mm.	Trace
Medium sand	$\frac{1}{4}$ to $\frac{1}{2}$ mm.	Trace
Fine sand	$\frac{1}{8}$ to $\frac{1}{4}$ mm.	6 per cent
Very fine sand	$\frac{1}{16}$ to $\frac{1}{8}$ mm.	23 per cent
Silt	$\frac{1}{256}$ to $\frac{1}{16}$ mm.	60 per cent
Clay size	Smaller than $\frac{1}{256}$ mm.	11 per cent
Total		100 per cent

Active channels, crests of natural levees, beaches, and bars, where winnowing of sediments prevails, tend to have the coarsest sediments. Deposits of active channels consist chiefly of fine sands with great variation in grain size and degree of sorting. The finer sediments are concentrated in the quiet water of lakes, marshes, stagnant bays, and quiet channels. Sorting is generally poor in deposits of high-water stages, and the coarsest and best sorted sediments are then in the pools. All areas of slack water have sediments ranging from the coarsest to the finest carried by the river.

Areas within range of active currents become floored with well-sorted fine sands from which any finer materials have been removed by currents. Places of slack water and bottoms of comparatively stagnant pools are covered with soft and poorly sorted oozes.

Deposits of natural levees range from fine sands to clayey silts. The coarser sediments dominate on and just inside the crests, and the average sediments range from coarse silts to very fine sands.

Deposition is relatively slow in marshes, and deposits consist of organic matter, mostly of plant origin, and fine silts and clays brought in when flood waters spread over the marshes. Areas protected against floods receive little detrital matter, and the sediments consist of nearly pure organic materials containing marsh gas and hydrogen sulphide. Sediments of small lakes and bays protected from flood water are fine clayey oozes of which some parts are gels that can stand unsupported to 6 or 8 in. high and hang together even when subjected to considerable jarring. The range in color is from metallic bronze to black. Thickness ranges from 2 to 3 ft., and the oozes are generally underlain by a very dark greenish or bluish clay. Similar oozes may accumulate in stagnant channels and some open bays, but most deposits of the latter range from poorly sorted silty clay to well-sorted very fine sand.

Beaches that margin the lower delta contain the coarsest and best sorted sediments of the entire delta, exclusive of the dune sands associated with and derived from them. Pebbles are occasional but scarce. Sources are not known.

Sediments deposited in open bays beyond the passes are distributed by waves and currents of the Gulf with much responsibility to currents of the river. These sediments have considerable variation in grain size and degree of sorting in both lateral and vertical distribution.

Russell and Russell describe the delta as "essentially a thick lens, with a relatively thin non-marine inner portion, a correspondingly thin outer marine portion, and a thick central zone composed of alternatingly marine and continental sediments." It is also a lens in direction parallel to the coast.

The fresh waters of the river flowing into and mingling with the waters of the Gulf create an environment of biologic instability from which there arise faunally desert conditions for considerable distances outward from mouths of distributaries. Neither normal marine organisms nor those of fresh water find the environment a congenial one.

The extents of the components of deltas depend upon the several variables of character and supply of sediments, extent of delta front, depth of water in which a delta is being built, velocity and volume of the stream, vigor of waves and currents, and movement of sea level.

Large supply and coarse sediments lead to rapid advance of a delta and wide subaerial and subaqueous plains, whereas, if materials are fine-grained and supply is small, the waves and currents have little difficulty in spreading the sediments so that most of them appear in the foreset and bottomset beds. Deep water gives large development of foreset beds, and vigorous wave and current action spreads the sediments widely, so that much becomes bottomset deposits. Moderate depth coupled with wide extent of delta front and a slowly rising sea level produces a large subaqueous plain, whereas weak waves and currents coupled with abundant sediments and stationary or falling sea level favor the development of a large subaerial plain.

As a delta is a transitional marine and continental environment, the deposits of each interlens with the other. On the marine side the sediments pass into others that are wholly marine, and on the other side into those of the valley-flat (Fig. 19).

The individual characteristics of deltas may all be found in other environments, but the associations and the larger picture can be seen only in a delta deposit. It is needless to state that a delta deposit cannot be proved by examination of a few exposures over a small area, unless the delta deposit is equally small.

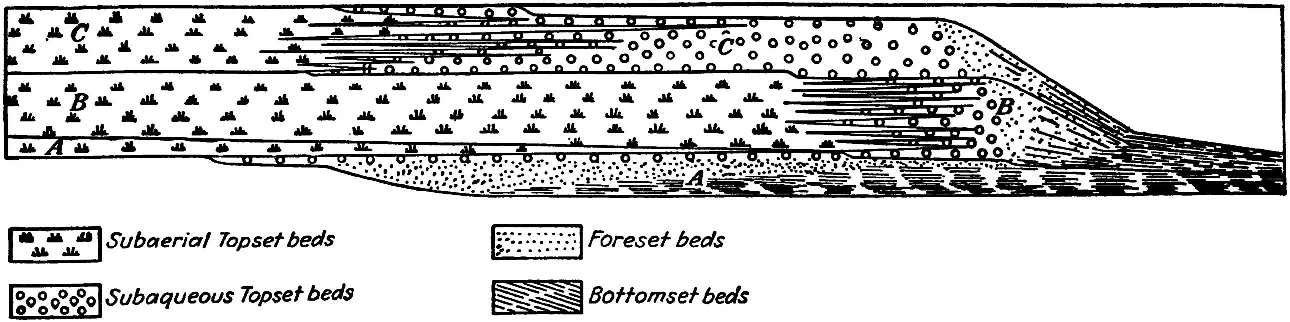

FIG. 19. Diagram showing relations between delta building and subsidence. *A*. Delta built into water with stationary sea level. Basin deeper than base level of deposition. Deposition produces shallowing. *B*. Sea level rising. Subsidence balanced by deposition. Delta built upward rather than outward. Subsidence intermittent. *C*. Subsidence at a faster rate with larger extent of subaqueous topset deposits. (*After J. Barrell, Bull. Geol. Soc Am.*, 23 (1912), 399.)

The Delta Cycle (*Barrell,* 1912)

The development of the three components of a delta is related to the physiographic age of the region drained. A region in the stage of physiographic youth and maturity supplies more sediments than the waves and currents of the body of water, into which a stream empties, can carry away. Moreover, a stream is likely to enter the sea at the head of a bay in which the water is not deep and the vigor of the waves and currents is at a minimum. A subaerial plain rapidly develops. As the delta continues to advance, it progresses into deeper water; the vigor of waves and currents increases; they become more competent to transport; and more and more sediments are added to the foreset and bottomset parts of the delta. The attack of waves retards extension of the subaerial plain and increases the area of the subaqueous plain. During this sequence of events, distributive provinces of the sediments have been eroded and have become unable to supply as large a volume of sediments as previously. Ultimately, with decrease in supply, a balance is attained, and waves and currents can transport and spread all that is supplied. Thence, farther in the cycle, as supply continues to decrease, there is a gradual increase of transporting ability of waves and currents over that needed to move the sediments supplied. This excess energy attacks deposits already made and ultimately may erode all to the base level of erosion, transporting the sediments outward to make marine deposits above and beyond the bottomset deposits of the delta. These deposits may be *coarser* than the underlying sediments despite the fact that they *are made in the old-age stage of the cycle of erosion on the land.* Topset beds of the subaerial and subaqueous plains and goodly parts of the foreset beds may be removed to the base level of erosion. The surface made through removal of the sediments is an eroded one, but there *has been no diastrophism of any kind.* This sequence of events with the time involved is the delta cycle.

The foregoing considerations assume no movement of the regions of supply or of sea level during the cycle. If, however, the distributive provinces were elevated more or less constantly, the result would be merely to continue to extend the delta. If elevation of distributive provinces was postponed until the cycle was nearing completion, there would begin anew the building of a new delta over the remains of the old, and the two deposits would be separated by a disconformity.

If sea level should more or less slowly rise as the delta was developing and if there was no increase in the supply of sediments, there would be little or no increase in the area of the subaerial plain. The subaqueous plain would increase in extent, and the topset beds might attain a thickness equal to the rise of sea level. If distributive areas should rise at the same time, the supply of sediments might be sufficient to maintain a subaerial plain. If rise of sea level was intermittent, but each rise so great that delta building could not keep pace, there would be advance of the sea with each rise and erosion of the delta surface on the shoreward side and deposition of marine sediments on the seaward side. These marine sediments would rest on an eroded surface. Later the delta might readvance, and this sequence might be repeated several times. Each marine unit would have a disconformity at its base. If movement of sea level was oscillating.

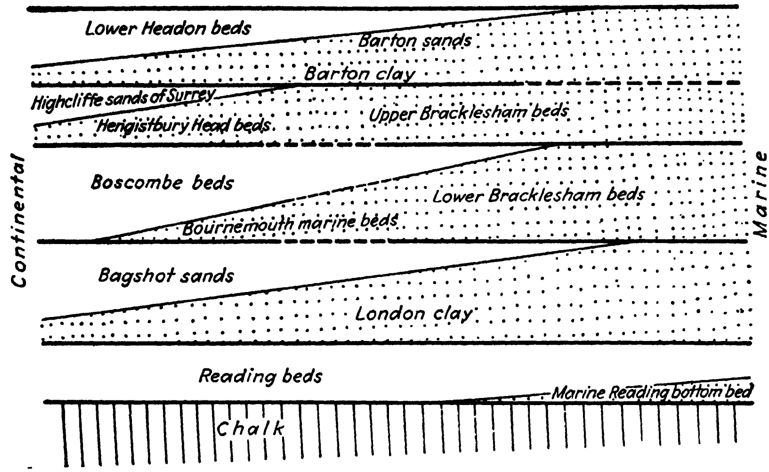

FIG. 20. Diagram showing interfingering of marine and continental deposits in deltas. The diagram also shows cycles of sedimentation. Marine strata shown on right, continental on left. The lines bounding the marine units show migrations of the shore. Vertical sections show geograms at the places indicated. Each cycle begins and terminates in a "ravinement." The diagram illustrates the conditions in the Hampshire Basin of England. (*After L. D. Stamp, On cycles of sedimentation in the Eocene strata of the Anglo-Franco-Belgian Basin, Geol. Mag.*, 58 (1921), 153.)

each rise would mean a marine deposit over the delta, each fall a delta deposit over the marine (Fig. 20).

In general, it may be expected that the delta cycle would be complex rather than simple and that a section through a delta deposit would show several disconformities, several marine units, and several units of strictly delta deposition, of which some would be complete from bottomset to topset beds and others incomplete, showing only a part of the delta sequence.

Extent of Delta Deposits

Little is actually known of the extents of submerged portions of deltas, but subaerial parts may be measured. The area of the subaerial plain of the Ganges and Bramaputra delta is placed at 50,000 square miles. That of the Nile has about the same area. The delta of the Hwang Ho has a greater extent, and the combined areas of the subaerial delta plains of the Rhine, Rhone, Po, Volga, Indus, Mississippi, Amazon, Orinoco, Colorado, Yukon, and others measure several hundred thousand square miles, perhaps more than ½ million square miles. This may be doubled or even tripled if the subaqueous parts are included, so that the total area of the earth's surface now receiving delta deposits may total 1 to 2 million square miles. Each period of geologic time may have had delta area of somewhat equal magnitude, but many of these ancient deltas are probably buried beneath the sediments of the continental shelf and continental slope.

It is thought that the oldest delta deposits are in the Pre-Cambrian, as parts of the Huronian strata resemble delta sediments, but there is no decisive proof, owing to the impossibility of proving the association of marine and continental deposits. It is probable that the formations of the Appalachian Region that form the upper part of the Ordovician and the lower part of the Silurian—the Queenston-Medina and the Juniata-Tuscarora—represent, at least in part, deposits of the delta environment. The Devonian Catskill "formation" of New York and Pennsylvania is generally considered to be of deltaic origin and to represent deposits made on the subaerial plain. These sediments grade westward into others which represent other components of the delta. The New Providence shale (Mississippian) of Kentucky is of deltaic origin, and the Chester formations as exposed in western Kentucky are thought to represent a dovetailing of marine and deltaic deposits. This is postulated to have been due to intermittent rise of sea level. The "Coal Measures" of Indiana, Illinois, and western Kentucky seem best interpreted as having been deposited under progressive but intermittent rise of sea level with occasional movement in the reverse direction. The deltaic materials consist largely of yellow sandstones of which some are highly micaceous; yellow, gray, blue, and black shales; and beds of coal. Units of marine origin, as shown by the fossils, interrupt the sequence of continental sediments. These in many cases are limestones and in contrast to the clastic sediments of deltaic deposition have rather wide distribution. They record extensive invasions of marine waters over a delta plain that seems largely to have been subaerial. The "Coal Measures" of Kansas, northern Oklahoma, and Missouri were also largely deposited in the delta environment, but they differ from those of Illinois, Indiana, and western Kentucky in containing many more marine units, thus indicating more frequent invasions of the sea. Moreover, as most of the marine units have considerable thickness, it is probable that the invasions endured for longer times. Rise of sea level was also progressive but intermittent, and there were also recessions, as channels of considerable depth are known to have been cut in previously deposited sediments. The deltaic deposits consist of sandstones, shales, and coal beds. The marine units are usually limestones and shales and, rarely, sandstones. These dovetail in intricate manner showing that the marine invasions were probably of different magnitudes. In southern and southeastern Oklahoma and adjacent parts of Arkansas the "Coal Measures" are also of deltaic origin, but here they seem to be largely of the subaerial environment. Considerable parts of the Permian "Red Beds" of Kansas and Oklahoma are considered to be of deltaic origin with deposition taking place largely on the subaerial plain under dry conditions with no possibility of reduction of the iron to which the redness is due. The great pile of "Coal Measure" sediments of north central Texas seems in large part to have been deposited on the subaerial plain of a delta, but some parts of the section seem to be of subaqueous deposition, and there are also marine units. The thick Pennsylvanian sediments of the Appalachian Region are considered to have been largely deposited on the subaerial plain of the delta environment. Schuchert and others have suggested that there were three great deltas, of which one centered about the present region of Pottsville, Pennsylvania, a second in the region of the present Kanawha Valley of West Virginia, and a third in the region of the present Cahaba Valley of Alabama. There probably were others.

The Cretaceous strata of the western part of Montana and Wyoming consist in large part, if not entirely, of delta deposits made by rivers draining regions on the west and flowing into the Cretaceous sea to the east. The sediments were largely sands, silts, and clays, of which the primary colors seem to have been gray to black. Coal beds are more or less occasional and have best development on the western border. It is thought that sediments of all delta components may be identified (Bowen). Fine dark shales containing marine fossils are referred to bottomset origin; cross-laminated sandstones and shales containing marine fossils are referred to foreset deposition; and shales, cross-laminated sandstones, and coals represent the environment of the subaerial and subaqueous plains. Berry has suggested that the Cretaceous Tuscaloosa formation of Alabama and Mississippi is of deltaic origin.

Delta deposits are apparently common in the Tertiary of the western coast of the United States. The Ione formation of California has been referred to deposition in the environment of a compound delta made by many streams draining areas to the east. The Pico formation, 3,600 meters thick, of the Los Angeles Basin seems to be of deltaic origin (Cartwright). The deposits are noteworthy in containing units of gravels of which the average diameter of particles is about 2.5 cm., but ranges to about 30 cm. These are angular to round and evidently were not transported far. Beds of gravel are up to 1.5 or more meters thick.

Delta deposits are probably as common on other continents as on North America. The Weald of the Cretaceous of southeast England has been referred to origin in the delta environment and so has also the Jurassic Estuarine series of Yorkshire. The Pennsylvanian Millstone Grit has been considered as representing a delta made by a river flowing from the north. It is probable that considerable parts of the "Coal Measures" of Europe and the other continents are of deltaic origin. Parts of the Flysch and the Molasse of the Alps were probably deposited in the environment of a delta. Parts of the Tertiary of the London, Paris, and Hampshire basins represent delta deposits. The Tertiary deposits of the Irrawaddy Basin of Burma have been interpreted as deposits of a delta that filled an ancient Burmese gulf under conditions of a fluctuating sea level (Stamp).

THE ENVIRONMENT OF THE MARGINAL LAGOON

A marginal lagoon may begin its development in bodies of shallow water of which the bottoms are above the profile of equilibrium and must be eroded to that level. Incoming waves must break some distance from the shore and a barrier beach results with a lagoon behind it. Lagoons may also be formed through the building of bars across bays or over a peneplaned or delta area that is being submerged by the sea, or the bar may consist of an elevated area on the land submerged. More or less large basins may be formed in the last way which are connected with the main body by narrow openings (Woolnough). Openings into the main body are generally known as inlets. In parts of the St. Lawrence Region a lagoon is designated a *barachois* and the inlet a *tickle*. The term tickle, also, applies to any narrow passage connecting larger bodies of water. The statements that follow apply to lagoons connected with the sea.

There is a lagoon cycle. The lagoon receives sediments from the land, the atmosphere, and the sea. The streams flowing into the lagoon bring those from the land. Those brought from the sea enter through inlets and by wash of waves over bar or barrier. Dust is dropped from the atmosphere. Plants and animals make deposits in lagoons. These sediments build up the bottom; ultimately the lagoon becomes a swamp; and it may be built above sea level. Waves beat against the seaward side of the barrier and it is eroded. Waves and winds carry sediments across to the other side; so as it is eroded on the seaward side, it is extended on the landward through encroachment on the lagoon. Ultimately the barrier or bar may migrate entirely across the area of a lagoon. In this migration the deposits are eroded to the base level of erosion. As the base of the bar and the base of the lagoon sediments were deposited on a surface above the profile of equilibrium, it is likely that nothing of either the lagoon sediments or the sediments of the barrier would remain. The sequence of events and the time required to accomplish this are the lagoon cycle. If sea level should rise while the cycle is in progress, the sediments of both bar and lagoon might be brought below the base level of erosion, and thus some of each would be preserved. It should be kept in mind that lagoonal sediments are margined on the seaward side by sediments of a barrier.

The waters of lagoons range from fresh through brackish to very salty. At times of rising tides large quantities of salt water may enter lagoons. If inlets are wide and the quantity of fresh water entering from streams is not extremely large, the waters in a lagoon may be brackish. If the quantity of fresh water entering a lagoon is very small and the inlets are wide and deep, the waters in the lagoon are salty. If no water flows from the land and the region around a lagoon is arid, its waters become highly concentrated and evaporites may be deposited.

Organisms flourish in lagoons if salinity and turbidity are not prohibitive. Typically marine organisms dwell where the waters are normally saline. Dwarfed forms of normal marine organisms may occur if waters are brackish, and this probably is also the case if circulation is poor and hydrogen sulphide is formed. Fresh-water organisms may live in part or all of a lagoon. Conditions may be such that a lagoon is unpopulated by bottom-dwelling forms or even planktonic forms, but large numbers of planktonic and even nektonic forms may be brought in when storm or high tide waters flow over the bar. These organisms die and sink to the bottom to be preserved, as no scavengers are present to destroy them. The quiet waters of lagoons may encourage the growth of much floating vegetation, and, if conditions permit, this invites the presence of an animal assemblage adapted to the conditions. The rather general instability of the environmental conditions often leads to wholesale killing, and both plants and animals may be floated in from the sea, and if the waters are not of normal salinity, these are killed. The conditions are favorable for the entombment of much organic matter.

The sediments deposited in marginal lagoons range from typical marine sands, silts, clays, and calcareous sediments in the lower part of a section to peats at the top. Streams introduce sands, silts, clays, and perhaps gravel. Currents through inlets may bring in sands, silts, clays, and calcareous matter. Carbonates are precipitated organically by plants and animals and chemically by escape of carbon dioxide. Some calcareous sediments may be very fine-grained. Many lagoons have poor circulation. Bottom waters and sediments of such lagoons may become populated with sulphate-reducing bacteria with the production of hydrogen sulphide and ultimately the black monosulphide of iron. Black shales containing pyrite or marcasite and excellently preserved delicate organic structures are the final product. Evaporites are deposited in lagoons that do not have enough inflow from the land to care for the water lost through evaporation. The deficiency is supplied by inflow from the sea, and ultimately the concentration becomes so great that salts are precipitated. If the surrounding land is low, either peneplaned or of a plain construction, the clastic sediments brought to a lagoon may be of very small quantity and the particles of small dimension. Black muds rich in organic matter would result, the organic matter either coming from the land in the form of such small particles as spores or spore cases, brought from the sea in the form of plankton drifted in through inlets, washed over bars by high waves, or it may have grown in the lagoon.

The stratification of sediments deposited in marginal lagoons should be regular and even because of the general quietness of the waters. It should be more or less gently inclined toward low parts of a lagoon from the shores of the barrier and the land. Cross-lamination may be present in and about the inlets, in the mouths of streams, and in the sands of the barrier. The general shallowness makes it probable that wave ripple mark and wave markings would be common, and current ripple mark may be present along the lines of stream and inlet flow. Tracks, trails, and bubble markings may be common. Colors of sediments range from light to black. Color due to ferric oxide of iron should not be present except, perhaps, in connection with evaporites. The commonness of organic matter favors reduction of ferric iron.

Lagoonal deposits are never large and are only marginal to marine or lacustrine sediments. The fact that most lagoons are only temporary arrangements and are destroyed on completion of the lagoon cycle does not favor extensive occurrence of lagoonal deposits in the geologic column.

Lagoonal origin has been suggested for several deposits of the geologic column, but, except for coral-reef lagoons, there is no good evidence for them. The Birdseye limestone of New York

(Chazy) has been referred by some to lagoonal origin, but acceptable proof has not been produced. Bass has suggested that the "shoestring sands" in the Pennsylvanian of Greenwood and Butler counties of Kansas are buried offshore bars.

A lagoonal origin implies that a barrier of some kind existed, and proof of lagoonal origin requires that not only must the deposits of the lagoon be present but also those of the barrier. Stratification must accord with the boundaries of the lagoon. It is needless to state that lagoonal deposits merge into other deposits and cannot have wide extent. Fossil contents may be quite different from those of the adjacent waters.

The extensive shallow epicontinental seas of the Paleozoic and Cretaceous of North America presented ideal conditions for the development of marginal lagoons and barriers, and the frequent rising of sea level should have made their preservation possible. No really proved example has been found. This may rest upon misinterpretation.

The Environment of the Estuary

An estuary is that part of the lower course of a river affected by tides, and its extent depends upon the tidal range. Rising tides hold back the stream waters, and the quantity thus held back depends upon the height of tides. When tides reverse, the stored energy forces the waters out with considerable velocity. For an estuary to be of much importance the tidal range should be high. Marine waters flow upstream at high tide, often at the bottom, and fresh waters may be flowing outward at the same time on top. The estuarine environment passes seaward into the shallow-water marine and upstream into the fluvial. The limits of contact of fresh and salt waters fluctuate with the tides. Many streams with deltas have estuarine deposits in the channels.

Outlines of estuaries are more or less triangular. They attain greatest area under conditions of coastal submergence when they are most enlarged and extend farthest upstream. Owing to the triangular and often funnellike shapes of estuaries, they often have tides larger than on open shores.

Important existing estuaries are the upper part of the Gulf of St. Lawrence and lower part of the St. Lawrence River, upper part of the Bay of Fundy, the Chesapeake and Delaware bays, and estuaries of the La Plata, Thames, and Severn. Although there are many estuaries, they form but a small part of the total extents of the sedimentary environments.

There is much turbulence of waters in estuaries in the twice daily contest of tides and stream waters. This conflict checks velocity and causes much deposition of mud and sand about the margins and over the bottom. As waters flow out of estuaries with great velocity at times of low tide, deposits are removed from the channels, but those on the margins, which may be in the form of more or less broad and long mud and sand flats, may remain.

The twice daily replacement of salt water by fresh water and the corresponding replacement of fresh water by salt water produce an extremely unstable environment to which it is difficult, if not impossible, for animals to become adapted. Moreover, there is oftentimes great turbidity, and outgoing currents are frequently extremely strong. Owing to the several unfavorable conditions, estuaries ordinarily do not have a large population of benthonic animals and plants, and those that do live there are likely to be smaller than normal. The organisms of estuaries are apt to be unlike in different parts. Fresh-water forms live above the influence of salt water, and salt-water forms below the influence of fresh water. It does not follow, however, that shells of normal marine organisms may not be found nearly to the head of an estuary and the shells of fresh-water organisms well toward the seaward side, as the ascending and descending currents are strong and may be competent to transport fairly large shells landward or seaward.

The sediments of estuaries are derived from streams and the sea, and both may be rather thoroughly mixed together. In the estuary of the Severn River the sediments consist of clays, silts, and sands. These contain organic matter in the form of coccoliths and coccospheres similar to those of the adjacent Atlantic, tests of foraminifera, fragments of coelenterates, minute spines and fragments of echinoderms, tests of diatoms, and occasional entire but generally broken shells of other invertebrates (Sollas).

The stratification of estuarine deposits tends to be irregular, and some cross-laminated units may be bounded by planes that are not parallel. Current ripple mark and cross-lamination have decided variations in slope and direction, and bedding planes may indicate decided turbulence and choppiness of waters. On the other hand, the mud flats on the borders of an estuary may have excellent stratification. Upstream the sediments pass into typical fresh-water channel deposits and laterally into deposits consisting of mud-cracked clays and silts, occasional sands, and perhaps peats. Reliance for determination of an estuarine deposit can be placed on no single characteristic, but the entire association of sediments should be present, and these should have an outline characteristic of an estuary.

Estuarine deposits must have some distribution in the geologic column. Grabau suggested that black graptolitic shales of the Utica type were deposited under estuarine conditions. This carries with it the logical conclusion that the estuary was the place of killing of the graptolites and not their normal habitat. The waters of deposition were not the places of life, but of death. Although the writer believes that the places of occurrence of abundant graptolites record places of death, it is seriously doubted if many of these were in estuaries (see black shales). Few if any estuarine deposits have been clearly established. They must be present, however.

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MARINE ENVIRONMENTS

INTRODUCTION

Marine environments have been placed in the three classes of neritic or shallow-water bottoms; bathyal, the bottoms intermediate in depth to the deep sea, from 100 to 1,000 fathoms; and abyssal, or the bottom of the deep sea below 1,000 fathoms. Sediments deposited in the marine environments are of terrigenous, organic, volcanic, meteoric, and perhaps magmatic sources. Most of the sediments are of terrigenous and organic origin, but there are very important contributions of volcanic sediments (Fig. 21).

The bottom of the sea is populated in various degrees of density by marine animals and plants, each adapted to the various factors that control distribution of aquatic marine life. Bottoms that do not have congenial conditions may be like deserts on the land, whereas bottoms that have an abundance of food and other favorable factors may support populations of extraordinary density. The numbers of scavenger organisms and predators are rather closely adjusted to the supply of food. If rates of deposition are slow, organic matter is not buried, is accessible to scavenger organisms, and is eaten. Shells are broken and after repeated passes through intestinal tracts may almost completely disappear. A bottom supporting an abundant population may yield few macroscopic organic sediments. On the other hand, a bottom with small population but fairly large deposition of sands and muds so as to provide frequent burial, may produce sediments filled with fossils. The generalization is made that *the number of fossils in marine sedimentary units bears little relation to the abundance of life on bottoms of accumulation* (Wepfer).

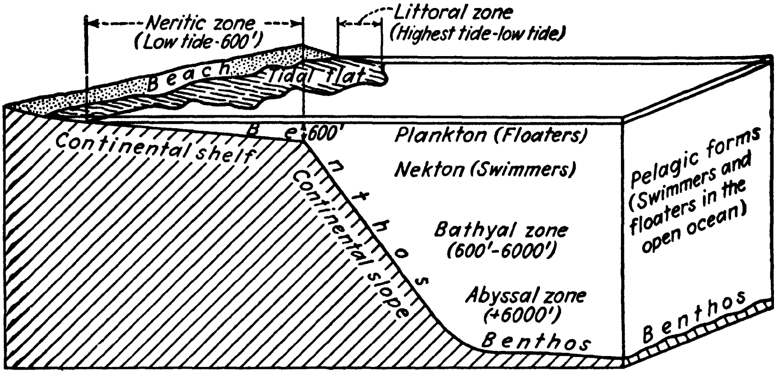

FIG. 21. Diagram showing the several marine environments. (*After Twenhofel and Shrock, Invertebrate paleontology* (1935), 12.)

THE NERITIC ENVIRONMENT (SHEPARD, STETSON)

Introduction

The neritic environment is more or less generally confined to the continental shelf and the epicontinental seas, as the Baltic Sea and Hudson Bay. The existing area is estimated to be between 10 and 15 million square miles. The neritic zone of the sea bottom is narrow about some coasts, such as the western coast of South America, and it is generally narrow about oceanic islands. Coasts that have been stable for a long time, as the eastern coast of North America, or have recently been submerged, are apt to have wide neritic bottoms.

If sea level rises, the outer part of the neritic environment becomes a part of the bathyal, and the bathyal loses its outer part to the abyssal. The last is not important, but the change from neritic to bathyal may take the bottom from the zone of light to that of darkness. A part of the land is added to the neritic, which may be very large if low lands are submerged, as was so frequently the case in the Paleozoic when the area of the neritic environment may have been two or more times what it is at present. Walther advanced the view that there was no deep sea until after the Paleozoic. If such was the case, the neritic sediments of the Paleozoic may have covered extremely great areas. It is thought that Walther's suggestion is not correct.

The neritic environment has several expressions depending upon whether it margins the open sea or an epicontinental sea, and the entire area of an epicontinental sea may belong to this environment. The environment may margin the mainland; coral, volcanic, or other islands; or lie upon a submerged area surrounded by deeper waters. Coral and volcanic islands are not so likely in epicontinental seas, but other islands and submerged areas surrounded by deeper waters may be present. Strong waves and currents prevail in the open sea, and tides and tidal currents are invariably present. This is the typical neritic environment. Waves and currents in epicontinental seas may average less strong than in the open sea, and tides may be small or wanting, as in the Baltic Sea. On the other hand, the coastal configuration may be such that the tides in epicontinental seas are particularly strong and high, as they are in Hudson Bay and the Bay of Fundy.

The waters of the neritic environment of the open sea are of normal salinity, whereas those of the epicontinental seas may range from *nearly or entirely fresh* to excessively salty and may change in salinity from time to time owing to restriction of connection with the open sea or to large inflow of fresh waters from streams. It is suggested that this was the case for the Cambrian seas of the upper Mississippi Valley. Such changes are expressed in elimination of previous organisms, so that fossiliferous strata are followed by those barren of fossils.

Floras and faunas tend to be more cosmopolitan in the open sea than in epicontinental seas, and in many cases the organisms of the latter are provincial. Submerged areas with deeper waters around them offer optimum conditions for bottom organisms, as there is little possibility of the entrance of much terrigenous sediment. It does not follow, however, that deposits over such places would contain an abundance of determinable organic remains, as scavengers adjusted to the food supply would be likely to destroy organic remains almost as rapidly as formed, and the deposits might be without macroscopic fossils.

Neritic environments marginal to large land areas receive greater or less volumes of terrigenous sediments that in volumes, but not in coarseness, express something of the stage in the physiographic cycle; and in their coarseness, rounding, and mineral contents something of the distance of transportation. The neritic environments about coral islands should receive few terrigenous sediments, but about volcanic islands these and volcanic sediments should be abundant. As coral and volcanic islands have steep slopes, there is likelihood of high initial inclinations, and there may be considerable slumping.

The organisms and kinds of calcareous sediments about coral islands are highly variable. The sediments and organisms of the lagoons are unlike those on the surface of the barrier, outside the barrier, and in the channels through the barrier, and each of these three environments is unlike the others.

Processes of the Neritic Environment

The sedimentary processes of the neritic environment are physical, chemical, and organic, each tending to be dominant at different places and at different times at each place. It seems probable that physical processes have equal, if not superior importance to chemical processes as essentially the whole of neritic bottoms is within the range of wave and current action of which the intensity is more or less inversely proportional to depth of water. Tides and tidal currents affect all open neritic water and waves and currents have great variations in intensity and directions of movement. The sediments of the shallower parts of the environment are repeatedly shifted, worn, and rounded, and shells, even when occupied by living animals, are often broken. Broken shells are repaired in some cases and thus remain as evidence of the vigor of the physical processes. However, it is possible for shells to be broken by predator organisms and to be later repaired, so that healed shells do not prove wave activity. The fact of wave activity affecting neritic bottoms indicates that these are nearing, and perhaps have reached, the level of the base level of deposition, so that in many instances the sediments have but temporary deposition.

Streams enter the sea by way of the neritic environment and leave most of the sediments carried by them upon neritic bottoms, where deposition may be permanent or temporary, depending upon the position of the bottom with respect to the base level of deposition. If the surface of the deposits is above that level, they are later shifted to deeper bottoms. Deposition is accomplished by checking of current velocity and action of electrolytes and colloids of sign opposite from those of the sediments.

Weak tides favor quiet water in deep holes of the bottoms, in bays and sounds, and in shoreward parts of shallow waters that extend long distances outward from shore. Limitation of agitation retards entrance of oxygen and escape of carbon dioxide and favors production of hydrogen sulphide.

Rates of deposition probably range from nearly 1 meter per day at some places to an equal thickness at other places in many thousand years.

Bottoms that have attained the base level of deposition receive no permanent deposits. The sediments that cover a bottom of this depth are repeatedly washed, fine materials are removed, and coarse sediments remain to form the bottom.

Ripple marks of all kinds, bubble impressions, pit and mound structures, cross-lamination, and all other features due to physical processes that may form under water may be present in all neritic deposits.

Chemical processes proceed in the neritic environment wherever waters carrying different substances in solution mingle, as at the mouths of rivers; where there is much decay of organic matter, as is the case over numerous areas of the bottom; where there are changes in temperature leading to variations in the carbon dioxide content, as is the case where cold waters of the bottom rise to the surface in warm regions with consequent loss of carbon dioxide and precipitation of calcium and other carbonates; where green plants flourish and extract carbon dioxide and precipitate carbonates; where there is evaporation of sea water with possible precipitation of materials in solution; and where the conditions are such as to favor the formation of hydrogen sulphide, sulphuric acid, and the sulphides of iron. The last takes place on a considerable scale in nearly tideless epicontinental seas, as in the east Baltic Sea, the Black Sea, and in deep holes of the bottom.

It is probable that a greater abundance of life per unit area flourishes in the neritic environment than in any other place on earth. Bottoms that remain sufficiently solid and sufficiently free from sediments, so that these are not seriously disturbing factors, are densely carpeted with animals in several layers, and the overlying waters contain an equally dense population of planktonic and nectonic organisms. Neritic bottoms support mollusks, echinoderms, brachiopods, corals, and other aquatic organisms in countless billions. Each species is related to a particular phase of the environment. Certain bottoms because of shifting sediments, changing salinity, turbidity, temperature, or sludgelike character of sediments may be nearly barren of macroscopic life and thus in a sense are aqueous deserts. Bottoms to which access of oxygen is difficult, usually because of poor circulation, become populated with sulphate-reducing bacteria, and the bottom waters are so poisoned that they also are barren of macroscopic forms.

There is the possibility of devastating destruction of animal life in neritic environments, particularly in shallow waters and waters near land. The destruction of the tile fish in the North Atlantic in 1881–1882 is an example. A heavy rain in 1863 on the coast of Dorset, England, changed the salinity and killed

1,000 lobsters retained in traps, and a thaw on the same coast in 1865 melted so much snow and ice as to kill large numbers of *Octopus vulgaris.* Similar great destruction of fish and other marine animals by heavy rains on the coast of India has been described, and a great fall of rain in 1918 over the Great Barrier Reef of Australia killed all animals to the depth of 3 meters over some places on the reef. Brongersma-Sanders has called attention to numerous cases of the deaths of large numbers of fish. She ascribes the destruction to noxiousness of red water caused by the abundant presence of dinoflagellates which produce a poisonous substance which directly kills the fish. The blossoming of the dinoflagellates is thought by her to be due to upwelling water. The "red tide" which killed myriads of fish in Florida waters and made the beaches of the southwest coast unbearable for months is cited as an example. Galtsoff lists many examples of "red tides" and estimates that from 100 to 200 million pounds of fish were killed by the red tide off the southwest coast of Florida in 1947. Whatever the cause of the killing, there are no doubts of the occurrences in many parts of the world and all seem to be accompanied by redness of water. Such sudden killing of large numbers of fish and other animals may not permit scavengers to escape, makes the waters foul with decaying organic matter, produces waters filled with hydrogen sulphide, and may lead to burial of organic matter without danger of much destruction.

Distribution of Sediments

The coarsest sediments tend to be nearest the shore, but there are many exceptions, and on many shores fine sediments are nearest the shore and the coarser sediments are seaward. The thickest accumulations are adjacent to mouths of rivers. Shallow bottoms tend to have sediments frequently deposited and as frequently removed. This leads to reassemblage of sediments and vertical and horizontal variation. Organic deposits may have few macroscopic shell materials, if scavengers are sufficiently abundant to break and masticate shells, but shells may be abundant if the conditions are such as to produce early burial. The result is that a section along a line drawn from the coast to the outer edge of the neritic life zone may pass through every variety of clastic sediment deposited in those waters, and the same general sequence may be repeated a second time. The section may begin with gravels at the shore, pass into sands, then silts, and finally clays. Farther out calcareous sediments may be encountered, and there may be alternations of calcareous sediments with muds and sands. On the other hand, the sediments may be muds, thence pass to sands, to gravels, and then to these sediments in reverse order. A section across the bottom at another place may have calcareous sediments from the shore to the edge of the neritic zone. The extremes are great irregularity of distribution and equally great regularity. Sediments on the continental shelf of New England (Stetson, 1938) in the offshore direction are present in the following order: "relatively fine, well-sorted sand; coarse well-sorted sand; and poorly sorted silts and clays" and well-sorted sand at the break in slope. On the shelf south of Cape Hatteras "there is a near shore belt of fine, well-sorted sand, and from this point to the break in slope the sand is coarse and well sorted. The inorganic material is gradually replaced by shell fragments and calcareous oölites. Below the break in slope the Gulf Stream carries away all sediment." "There is no steady gradation from coarse to fine in an offshore direction" on the continental shelf of eastern United States.

Neritic Sediments

Neritic terrigenous sediments consist of gravels, sands, silts, and clays. Gravels and sands tend to be well sorted; many are long traveled; and quartz in some of its forms is a most common composing substance. Some gravels may have come but short distances. They may be composed of any kind of rock or mineral and be poorly to excellently sorted. Coarse sands of long travel tend to be well rounded; those of small dimension are usually angular. Muds have gray, blue, black, and intermediate colors. Red muds are deposited about the mouths of some tropical rivers, as the Amazon. If there is much organic matter deposited with the muds, the red color is not retained. Black muds have limited distribution and are deposited where circulation is poor. The blue, green, and gray muds have wide distribution. The green is thought to be due to glauconite, and the blue is probably referable to organic matter.

Chemical and organic sediments are abundant and consist of shells, algal and coralline matter, and some other substances. Most are calcareous, but some silica, calcium phosphate, iron oxide, and a few other substances are present. Appeal has been made to magmatic sources for some of the iron and silica, and it has been assumed that contributions from this source were more extensive in the early ages than later. This is possible. Calcium carbonate has extensive distribution over great areas of the neritic environment as well as over other parts of the ocean bottom. Much calcium carbonate is formed by plankton, mostly small plants and foraminifera. Where conditions are favorable the tests and shells of these organisms settle on neritic and other bottoms like a gentle rain.

The Marine Cycle

The marine cycle affects all shallow-water marine sediments. There are two phases. In one the shores have been made by submergence of a land area, and in the other by emergence of a sea bottom. The former is considered first.

In the early stages of a marine cycle initiated by submergence of a land area, the shores are irregular and may be high. Bottoms may descend abruptly from the shore. (If flat lands are submerged, this would not be true.) Sediments are apt to be supplied in excess of the disposing ability of the sea, and the bottoms may be built above levels that could not be maintained after the supply decreases. Headlands would be cut back and indentations filled. As the quantity of sediments supplied becomes less as sources are eroded to lower levels, any sediments that lie above the base level of marine erosion would ultimately be removed. The coast at the same time would be eroded inland and acquire a configuration adjusted to lithologic and structural response to wave attack. When the old age of the cycle was reached, all sediments above the base level of marine erosion would have been removed. Deposits in deep bays would escape removal to the extent that they lay below this level. Bay bars and lagoonal deposits behind them would also escape if they lay below this level.

Sediments derived from erosion of shores at any stage in the cycle would have moved along shores to bayheads and basinward to form offshore deposits on bottoms beneath the base level of erosion. These build the wave-built terrace which slopes basinward, the coarsest sediments at the top of the seaward slope, the progressively finer ones lower down. Each such deposit forms a time unit which slopes more or less gently in basinward direction. The wave-built terrace thus becomes overlain by the coarsest sediments that are carried outward, and these are underlain by sediments which are progressively finer outward. Each lithology forms a lithic unit which slopes basinward at a somewhat smaller angle than the time units. Thus, lithic units transect the time units.

If a shore is due to emergence of a gently sloping sea bottom, or submergence of a low plain, the new bottom is underlain by weak materials, and it is likely to have a slope more gentle than that of the profile of equilibrium and the new bottom is also likely to be above the base level of erosion. This condition would lead to the formation of barrier beaches with lagoons behind, and removal of the bottom to the base level of erosion would begin. This would progress until the entire bottom had been removed to this level with barrier and lagoonal sediments removed and the shore inland from the initial shore. It would have a configuration adjusted to lithology, structure, and the vigor of wave attack.

Extent of Neritic Deposits

Neritic deposits constitute most of the marine sedimentary rocks exposed on the lands. Sedimentary rocks cover about 45 million square miles of the land to an average thickness of about 1 mile, of which it is estimated that about 80 per cent, or 36 million square miles are underlain by sediments of neritic origin. Marine sediments of existing neritic environments are estimated to have a volume of 30 million cubic miles, thus giving a total volume to neritic sediments of 66 million cubic miles, which is more than twice the volume of all land above sea level (see Table 3, page 139).

THE BATHYAL ENVIRONMENT

The bathyal environment is that part of the sea bottom between the depths of 100 and 1,000 fathoms. This has an estimated area of 15 million square miles. The higher parts of the area receive a little light. The surface of this bottom has little slope except around coasts of coral and volcanic islands and fault coasts, where slopes may be steep. Bathyal conditions pass seaward into those of the abyss and landward into those of neritic bottoms.

Physical processes have limited affects on bathyal bottoms. Currents may be strong in upper parts and beneath straits connecting larger bodies. Volcanic and seismic waves may affect the bottom to great depths and Bramlette and Bradley found evidence of weak currents on the bottom of the North Atlantic to the depth of 1,955 meters. Clastic sediments on bathyal bottoms are dominantly fine-grained and are received through settling from suspension, through slumping or sliding down steep slopes from higher bottoms, and from currents moving downward from shallower bottoms. Many steep slopes are loci of seismic movement, and this may aid in dislodging sediments from higher to lower levels and in time may bring coarse materials to bathyal bottoms (Fairbridge). Deposition is thought to be slow and without important interruption.

Chemical processes are thought to be less varied than in the neritic environment. There is probably considerable solution of planktonic organic sediments settling from surface waters and precipitation of carbonates as carbon dioxide escapes at places of upwelling waters. Deep holes in the bottom may have poor circulation, and hydrogen sulphide and attendant products may form in these.

Organisms are probably responsible for many of the sediments on bathyal bottoms, and many of the organisms are plankton living in upper waters. The bottoms may also have a prolific population despite the fact that plant life is limited. The slow rate of deposition renders it probable that scavengers destroy much organic matter.

Stratification, if not destroyed by burrowing and mud-eating organisms, should be regular and uniform over great areas. It is probable, however, that the slow rate of deposition permits rather thorough working of sediments by organisms and mixing with those already deposited, and thus much stratification would not be shown. Where bathyal bottoms lie near shores it is probable that some lenticular distribution develops, but it is thought that lenses are on a larger scale than is common in the neritic environment. All strata should have initial inclination related to the slopes of the surface of deposition, and this may locally be high. Cross-lamination is likely to be limited and then largely over the upper part of bathyal bottoms. Current and wave ripple marks are also possible over this upper part, the latter rare. Current ripple mark may also form on deeper bottoms where coastal and bottom physiography produces deep currents.

Sediments deposited in the bathyal environment are sands, perhaps gravels, various muds, calcareous substances, glauconite, siliceous sediments, and various other more or less rare substances. Sands may be terrigenous or "volcanic" and are not common. Muds seem to be the most common sediments, and colors are red, gray, green, blue, and black. Most muds seem to contain considerable calcium carbonate. Some muds are of volcanic origin.

TABLE 2

Sediment	Area, square miles	Mean depth, fathoms	Depth limits, fathoms
Coral muds	2,236,800	740	1,820
Coral "sands"		176	
Volcanic muds	600,000	1,033	260 to 2,800
Volcanic "sands"		243	100 to 420
Green muds	850,000	513	100 to 1,270
Green sands		449	900
Red muds	400,000	623	120 to 1,200
Blue muds	14,500,000	1,421	125 to 2,800

Calcareous sediments consist largely of planktonic shells, and some are shells of benthos. Much calcareous foraminiferal ooze is deposited on bathyal bottoms, but as most belong in deeper waters, their consideration is deferred.

Siliceous sediments are represented by tests of diatoms, shells of radiolarians, and spicules of sponges. Diatoms make deposits of ooze on bottoms as shallow as 600 fathoms. Their tests may be present in any marine deposit.

The "Challenger" report (Murray and Renard) gives the areas and depths of the different sediments of the bathyal environment as shown in Table 2.

It has not been proved that sediments of bathyal origin are present in the exposed geologic column, but it is possible that sediments of the higher or shallower parts of this environment have been raised above sea level. Positive proof is lacking and proof is difficult, as the sediments of this part of the environment have no differences from those of the lower or outer parts of the neritic environment.

The volume of existing bathyal deposits is considered to be large, and it seems probable that few of them have ever been destroyed. It is also considered probable that interstratified with the bathyal sediments and covering some of the same area are neritic sediments deposited at times when sea level was lower and the lands more extensive. It is suggested that the volume of bathyal and interstratified sediments is of the order of magnitude of 40 to 50 million cubic miles.

THE ABYSSAL ENVIRONMENT

The bottom of the deep sea receives no sunlight, has pressures that slightly exceed 1 ton per square inch per mile of depth, and a temperature near freezing at all times. The water is without appreciable motion except when moved by volcanic eruptions or earthquakes. The area is about 115 million square miles and includes all depths below 1,000 fathoms.

No green plants and no plants of any kind that require sunlight live on the bottom of the deep sea. Animals dependent on plant food must live on that coming from the sunlit waters of the surface. Eyes become nonfunctional or modified to catch the feeble gleams emitted by the phosphorescent organs some animals have. The population is adjusted to the food supply and is smaller in numbers and variety than in shallower waters. Shells and bones of animals that live on abyssal bottoms are generally fragile, and many shells are transparent.

All five varieties of sediments are probably deposited on the bottom of the deep sea, but little is known relating to sediments of magmatic origin. Cosmic particles have little quantitative importance, but because of the slow rate of deposition of all kinds of sediment on deep-sea bottoms the cosmic particles attain an importance that is not known elsewhere. The relative quantitive importance of the other three kinds of sediments is not known.

Sediments of volcanic origin consist of ash and pumice in various degrees of alteration. After pumice is ejected it floats until waterlogged and undergoes some solution and decomposition. Some pumice is more or less deeply impregnated with manganese oxides when brought from the bottom, and doubtless some of this was acquired as it floated on the surface and sank from the surface to the bottom. Materials of volcanic origin are widely distributed on the sea bottom and are among the most common ingredients of deep-sea sediments.

Sediments of terrigenous origin attain the bottom of the deep sea by way of the air, water, ice, and in the roots and holdfasts of plants. Ice from polar regions floats into warmer water and drops its burden of sediments on melting. Particles carried by ice may have been derived from the land or the bottom of the sea, and dimensions may range to weights of tons. Fairly large particles may be rafted from shallow bottoms in the holdfasts of aquatic plants. These large particles introduce somewhat incongruous components into deep-sea sediments and make an association that has not been seen in the exposed geologic column. Terrigenous sediments brought by wind are finely divided and doubtless are composed of minerals of wide range, with quartz, limonite, and the clay minerals present in largest quantity. These sediments enter the sea at the surface and slowly settle to the bottom. In the passage downward they are probably leached of everything soluble. The quantity entering via the atmosphere is not known, but it is evidently large. Other terrigenous material is considered to be derived from that brought by the rivers to the sea and floated outward from the shore. The general assumption is that all finely divided sediments are flocculated and precipitated when they enter the sea. There is no proof that complete precipitation takes place, and it is considered not improbable that large quantities pass on to the deep sea.

Terrigenous and volcanic sediments at present are found over the entire bottom of the deep ocean, but in many places they are more or less masked by organic sediments. Where organic sediments are few, the terrigenous and volcanic sediments make the red clay. This covers 51.5 million square miles of the ocean bottom and ranges in distribution from the depth of 2,225 fathoms to the deepest known depth. Kuenen's estimate of the volume (1941) is 270 million cubic miles, the author's 83 million cubic miles.

Organic sediments are composed of shells and other hard parts of organisms. Many and probably most shells are those of plankton that dwelt in the upper sunlit waters of the surface;

others are shells of organisms that lived on the bottom. The latter do not seem particularly important, as the shells are fragile and the population is not large. The shells of plankton are small and delicate and evidently sink slowly, and some loss from solution may take place on all of them, and if the bottom is deep they may disappear completely before reaching bottom. Calcareous shells have largely disappeared at the depth of 2,500 fathoms, but siliceous shells attain deeper bottoms, with 4,000 fathoms about their maximum limit. Large areas are overlain by these organic sediments, and they are named after the most prominent shells. Two oozes are of calcareous composition and two are siliceous.[1]

The quantity of sediments on the deep-sea bottom composed of materials that can be considered of inorganic origin is certainly very large. Twenhofel has presented evidence that the total quantity is of the order of 120 million cubic miles. Kuenen (1946) found that the total mass of all deep-sea sediments based on different methods of calculation ranged from 2×10^8 to 10×10^8 and probably is from 6×10^8 to 8×10^8 and that the thickness of these sediments ranges from 1 to 5 km. and probably is around 3 km. The volume is several times that of all sediments on the lands.

Deep-sea sediments and also those of bathyal origin commonly contain crystals of phillipsite and nodules of manganese and phosphate. Manganese nodules seem confined largely to clays but are present in oozes. Resistant parts of upper-water pelagic animals are rather distinctive features. Among these are teeth of sharks and ear bones of whales. These are so common in some places as to be impressive as illustrated by a dredge made by the "Challenger" in the central Pacific from a depth of 2,385 fathoms. This yielded over 1,500 shark teeth and about 50 ear bones of whales.

Rates of deposition of sediments in the deep sea are not known, but the rates evidently are very slow as shown by the abundance of shark teeth, which on bottoms where rates of deposition are relatively rapid are by no means common. Analysis of all the evidence led to the suggestion by Twenhofel that a rate of deposition of 1 ft. in 87,000 years, or 2,900 years per centimeter represents a fair approximation. More recent estimates are those of Schott (corrected by Kuenen, 1941), 4,300 years per centimeter, Bramlette and Bradley about 300 years per centimeter, and Kuenen (1946, page 568), 2,000 years per centimeter. Other estimates have been made by Piggot and Urry. Schott in "Recent Marine Sediments" calculated average rates of deposition of different sediments in the equatorial Pacific as follows.

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[1] Planktonic organisms live in the upper waters everywhere, and they settle on shallow bottoms in as great if not greater numbers as on deep. There are so many other sediments on shallow bottoms, however, that the planktonic shells rarely do more than make a subordinate ingredient. Occasionally this is not the case.

Blue mud, 1.78 cm. per 1,000 years
Globigerina ooze, 1.2 cm. per 1,000 years
Red clay, 0.86 cm. per 1,000 years

In the southern Indian Ocean the average rate of deposition for Globigerina ooze was placed at 0.59 cm. per 1,000 years, diatom ooze, 0.54 cm. per 1,000 years. Revelle and Shepard for basins off the California Coast estimated the rate of deposition of clay and silt may be 25 cm. in 1,000 years. Kuenen's estimate of volumes and rates of deposition of deep-sea sediments is given in Table 3.

TABLE 3. VOLUME OF MARINE AND CONTINENTAL SEDIMENTS
(After Kuenen, 1946)

Sediment	Area, sq. miles	Mean depth, fathoms	Range, fathoms	Volume	Rate of deposition, years per cm.
Continental:*					
Sandstone				9,600,000	
Clays and shales				28,800,000	
Limestones				7,200,000	
Unweathered volcanics				2,400,000	
Deep-sea:					
Red clay	51,500,000		2,325 to deepest bottom	105,600,000	Pre-Cambrian, 6,000
Blue mud				26,400,000	Cambrian to Recent, 10,000
Globigerina ooze	49,500,000	2,000	400 to 2,925	12,000,000	5,000
Pteropod ooze	400,000	1,044			
Diatom ooze	10,800,000	1,477	600 to 1,075		
Radiolarian ooze	2,200,000	2,884	2,350 to 4,475		

* Continental sediments include all shallow-water marine sediments.

The extent of reworking by organisms of the deposits made on the bottom of the deep sea is unknown. The general scantiness of food suggests that the bottom materials are screened to the limit for any food they may contain, and thus original lamination or bedding of any kind would be destroyed. That some stratification is present is likely, but it probably would not be shown by parting planes. Sedimentary units may be assumed to maintain essentially uniform character, thickness, and horizontal position over large areas.

Deep-sea sediments may be summarized as follows: (*a*) They contain no large fragments derived from the land except as brought by ice, floating seaweeds, or trees, or slid down slopes from near-by lands. (*b*) They contain no indigenous plant materials. (*c*) They extend in essentially horizontal sheets of essentially uniform composition over immense areas. (*d*) Shells of shallow-water animals and shells of plant eaters should not be present. (*e*) Features due to wave and current action are rare or wanting. (*f*) The sediments contain resistant parts of pelagic organisms, which are not common in other sediments. (*g*) Indigenous shells of benthonic animals are thin and fragile as are also the more abundant shells of planktonic organisms.

Deep-sea Sediments in the Exposed Geologic Column

Walther made the suggestion that there are no deep-sea sediments earlier than the Mesozoic, and none is certainly known. Some geologists have referred the marine black shales to deep-sea origin, but there is little or no evidence for this view. The chalk of the Cretaceous system has also been referred to this origin, but this suggestion finds little support either in field relations or microscopic studies. Certain calcareous strata of the Alps of Europe have been postulated to have been deposited in the deep sea, but these may more easily be referred to shallow-water deposition. Radiolarites in the Alps and elsewhere have been referred to deposition in deep water, but positive proof has not been found to establish the suggestion. It has recently been suggested that Ordovician cherts of eastern New York were deposited in the Appalachian geosynclines in depths of abyssal magnitude. This was done on the basis of the small quantities of clastic materials in the cherts, the association of the cherts with graptolitic shales, and the presence in the cherts of the shells of fossil Radiolaria. The evidence comes a long way from being convincing, as every one of the characters named may be found in shallow-water deposits (Ruedemann).

It is stated that sediments of deep-sea origin are present on Malta in the Mediterranean, the Barbados and Christmas Islands in the east Indian Ocean. These are said to have been brought above sea level by local upheavals. The Dutch geologists (Molengraaf) have described deep-sea deposits on the islands of Borneo, Timor, and Rotti of the East Indies. The deposits, termed the Danau formation, cover some 40,000 square miles and belong to the Middle Mesozoic. The rocks are flint (hornstone) filled with closely packed radiolarians, and bright red argillaceous chert or cherty shale. The two varieties of rock gradually pass into each other. Nothing in the rocks can be interpreted as of shallow-water origin. Manganese is present as grains in the red shales, as grains and nodules in the red shales and cherts, and as slabs in the radiolarian flint. It is suggested that the chert represents clayey radiolarian ooze and the shale with radiolarian red clay. The Danau formation carries strong evidence of being of deep-sea origin.

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CHAPTER IV

ORIGIN OF INORGANIC SEDIMENTS

INTRODUCTION

There are five classes of sediments, which in probable order of importance are terrigenous, organic, volcanic, cosmic or meteoric, and magmatic. The sources of the terrigenous sediments are the rocks of the earth's crust. The organic sediments are derived from the supporting and protective structures of plants and animals and from products resulting from organic activities. Sediments derived from volcanic sources are considered third in quantitative importance. A small quantity of material comes from outer space to form the cosmic or meteoric sediments. It is thought that magmas entering the earth's crust may contribute waters with substances in solution. The union of these waters with surface waters is thought to precipitate materials from solution. The terrigenous, volcanic, cosmic, and magmatic sources of sediments are considered in this chapter in such detail as seems necessary. Organic sediments are discussed in Chap. V.

MAGMATIC SEDIMENTS

It is not possible to present concrete illustrations of magmatic sediments. That there are such seems highly probable, as magmas are thought to yield large quantities of highly heated liquids, mainly water, to surrounding rocks, and these must carry large quantities of matter in solution. Some of these liquids should reach the surface and mingle with surface waters. This should result in precipitation of particular types of sediments. Certain sedimentary rocks, as the Pre-Cambrian iron formations of the Lake Superior Region and the Franciscan cherts of California, have been postulated to be of this origin, but positive proof has not yet been presented. Waters which may be of magmatic origin are discharged into the sea from submarine springs near the volcano Santorin, situated on an island in the

Aegean Sea (Taliaferro). These waters contain dissolved iron carbonate.

COSMIC SEDIMENTS

Particles of cosmic or meteoric origin are constantly falling upon the surface of the earth, and it has been estimated that between 15 and 20 million daily enter the atmosphere. Although this number is large, its insignificance becomes apparent when it is realized that this number amounts to only one on each 10 to 13 square miles. Most are small. It is possible that the number was greater in the early history of the earth, and the meteorites may have been larger. According to a recent estimate, the quantity of cosmic material annually added to the earth is a minimum of about 100,000 long tons and it may be as much as 10,000,000 long tons. Accretion amounts to an average of about 50 grams per square mile annually, and it is estimated that at the present rate a layer of meteoric dust only 1 cm. thick has been deposited on the earth since its origin as a separate body in space (Watson). This is not a very rapid rate of deposition. Most of the cosmic particles are composed of nickel-iron.

The oxidation of small meteorites as they pass through the earth's atmosphere produces iron and nickel oxides, which are deposited over the earth's surface as dust in too small quantity to be detected. It is only the meteorites that escape decomposition in passage through the atmosphere that can possibly be recognized. There probably are many of these, and in the deep sea, where the rate of deposition is extremely slow, cosmic particles may rate high in the sediments as compared to places where other sediments are abundant. No meteorites have ever been found in the geologic column.

SEDIMENTS OF VOLCANIC ORIGIN

The clastic materials ejected by volcanoes are considered sediments if, when deposited from the atmosphere, they have cooled sufficiently so as not to cohere. The materials consist of the so-called ash, or the broken glass of lava froth; particles termed sands; and larger particles termed lapilli or cinders if small, and bombs if large. These particles are shot into the atmosphere with great force and thus placed in the regions of higher velocities of winds. The particles in the range from sands to bombs soon settle, but the "ash," with a range in the silt and clay sizes, is carried for longer or shorter distances from its sources, and there is little doubt that ash from a single eruption has encircled the earth. Ash has fallen everywhere and has been incorporated in all kinds of sediments.

Relatively pure deposits of "ash" are not uncommon. Many deposits are remote from volcanoes, as the ash beds of Kansas and Nebraska and the altered ash beds that form some bentonites. The "sands" and larger particles generally remain close to places of ejection, unless the eruption is subaqueous. Many large particles are porous and float readily and thus may be distributed over areas for hundreds of miles from sources. Slag, an artificial volcanic material from the smelters at Gary, Indiana, has been carried to all shores of Lake Michigan, and there is little doubt that it has extensive distribution in the bottom sediments of this lake.

The relative magnitude of volcanic sediments to others in the geologic column is not known. The actual quantity is large. An eruption that occurred in the Yukon Basin some time in the last 1,000 to 2,000 years is estimated to have spread 10 cubic miles of ash over a minimum area of 140,000 square miles. This gives an average thickness of a little more than 10 cm., which is not large. Katmai in Alaska in 1912 ejected 5 cubic miles of ash. Krakatoa in the Strait of Sunda in 1883 is thought to have produced a quantity fully as great, ranging from dust to large blocks, and the eruption of Tombora in the East Indies ejected between 20 and 50 cubic miles of materials. The Cretaceous of the Great Plains Region of the United States and Canada contains many cubic miles of altered volcanic "ash," or bentonite. Kuenen (1941) estimated that there are about 6 million cubic miles of unweathered volcanic products.

The figures given above are large, but compared to the many millions of cubic miles of organic and terrigenous sediments, they are small. The volcanic sediments really constitute but a small part of the sedimentary rocks of the geologic column.

TERRIGENOUS SEDIMENTS

Terrigenous sediments result from the destruction of rocks on the surface and in the outer parts of the earth's crust. These sediments depend for their detailed characteristics upon (*a*) the characters of the rocks from which the sediments were derived, (*b*) the methods by which the destruction was accomplished, (*c*) the environmental conditions of the distributive provinces, and (*d*) the distances of the distributive provinces from sites of deposition, which to some degree determine the time intervening between production and deposition. Sediments vary with the variety of rock destroyed, the methods of destruction, the environmental conditions of the places of production, and the distances and duration of transportation. A given rock produces a certain variety of sediments under one set of environmental conditions and another variety under another set.

SOURCES OF TERRIGENOUS SEDIMENTS

The source rocks may be igneous, sedimentary, and their metamorphic equivalents. The igneous rocks have been estimated to cover 25 per cent of the surface. The sedimentary rocks constitute only 5 per cent of the crust, but veneer 75 per cent of the surface. It seems not unlikely that some approximation to this ratio existed as far back as the beginning of the Paleozoic and that most of the sediments produced since that time have had sedimentary rocks as parents. It is not, however, so simple as the ratio indicates, as so many of the high parts of the surface are underlain by igneous rocks, and these high parts, because of position, produce large quantities of sediments.

Igneous rocks are composed of feldspars, quartz, amphiboles, pyroxenes, micas, magnetite and ilmenite, olivine, apatite, and many rarer substances of which some are extremely resistant to destruction and hence are likely to escape into sediments. If igneous rocks are destroyed by physical agencies, the sediments are composed of pieces of rock or mineral particles of the parent rock. If chemical methods prevail, the sediments are composed of derived products and minerals resistant to destruction. Leith and Mead have estimated that chemical destruction of the average igneous rocks yields sediments which, when lithified, consist of 82 per cent mudstones, 12 per cent sandstones, and 6 per cent limestones. Other ratios have been suggested.

If mudstones, sandstones, and limestones are physically destroyed, the products are merely pieces of the parents. Chemical destruction is not readily done, as these rocks in large parts are already the products of chemical decay, so that the products produced are not unlike the parents except that considerable soluble matter may have been removed.

Rock Destruction (Katamorphism)

Rock destruction may be accomplished by physical or chemical methods, and usually the two methods proceed more or less simultaneously. Physical destruction produces fragments that are minerals or pieces of the parents, whereas chemical destruction forms minerals that are different from the original rocks, the kinds of minerals formed being largely determined by the environmental conditions. The physical processes have been termed rock breaking or disintegration and the chemical, rock rotting or decomposition. Many of these processes are included under the term of weathering (Reiche).

ROCK BREAKING (DISINTEGRATION)

Rock breaking is accomplished by changes of temperature (insolation), frost action, rifting or wedging by growth of minerals or roots, diastrophism, undermining, abrasion by various agents, grinding, and impact.

Insolation. Rocks are not particularly good conductors of heat, and, if any part of a rock becomes sufficiently heated, it expands and may separate from parts of the rock that have not been heated. Many rocks when thrown into a fire break into fragments, some, like flint, seeming to explode, and, in some backward parts of the world, rocks are quarried by moving fires over the surface. Breaking in this way is evidently done in a natural way in prairie and forest fires (Blackwelder, 1927). The arching and breaking of concrete pavements on very hot days is a familiar phenomenon. It is generally supposed that as the outside of a rock becomes heated, the inside, owing to poor conductivity, remains cool, resulting in separation of the outside from the inside along cracks tangentially arranged. By evening, heat has penetrated to the inside, and during the night the outside cools and becomes too small for the inside. Radial cracks develop. It has been stated that the daily repetition of this performance ultimately brings about separation of spalls from the outer part along the radial and tangential cracks, resulting in exposure of new surfaces. This is one form of block disintegration and a phase of exfoliation. Doubt has been cast on the efficiency of insolation as a factor in breaking rocks into large fragments, and the results of a series of experiments in rock breaking conducted under the direction of Blackwelder led to the generalization that "insolation, or diurnal changes of temperature, is entirely inadequate to cause rock breaking" and that there is no "satisfactory evidence that insolation is ever the sole unaided cause of rock fracturing" (1933). Griggs subjected granite to artificial change of temperature in a 15-min. cycle ranging from 32 to 142°C. for 89,400 cycles, equivalent to 244 years, and no fracturing took place. This would seem conclusive for granite. However, there is some hesitancy on the part of the author in complete acceptance of Blackwelder's generalization and application of Griggs's experiments to all kinds of rocks as too many cases of breaking of artificial rocks from changes of temperature are thought to be known.

There is another phase of insolation that results in separation of the individual minerals. Each mineral has a particular coefficient of expansion, and in minerals other than isometric or amorphous, the coefficients of expansion may be different along the different crystallographic axes. The daily repeated expansion and contraction of the different minerals ultimately leads to complete separation. This is most extensively done in regions having the greatest extremes of temperature and should be more effective on dark- than on light-colored rocks. This is termed mineral or granular disintegration, and that it is effective may be shown by examination of such rocks as coarse-grained limestones that have lain on the surface for a long time. Block disintegration paves the way for granular disintegration, and insolation thus has as its final objective the separation of rocks into the individual composing mineral particles. In cases of igneous rocks this produces sediments that consist of feldspars, quartz, amphiboles, pyroxenes, micas, and rarer substances. Limestones yield calcite and dolomite. The sandstones may fall apart into the individual grains. Mudstones can do no more than break into small fragments, as it is essentially impossible for this rock to be separated by insolation into the individual mineral particles.

Insolation is limited to the surface, and thus its continuation requires removal of the broken products lest these cloak and protect the rocks below.

Frost Action. Freezing introduces the wedging action of ice. Water enters rocks in several ways, and if this is done before freezing weather, the rocks are certain to be broken when the water becomes ice. Water may be imprisoned in closed cavities at times of their formation in igneous and sedimentary rocks and minerals. These cavities burst when the water freezes. Beautiful crystals of quartz are known to have been broken in this way. Porous rocks are naturally affected more than nonporous ones. Permeable rocks with small pores suffer more than those with large ones if the rocks are in such positions that the waters may drain out. The results are that porous rocks after much thawing and freezing crumble into small pieces. In the case of sandstones like many of those of the Upper Cambrian of the upper Mississippi Valley a few freezings of the rocks when they are saturated with water suffice to crumble them to sand, whereas a granite, which has low porosity and little permeability, suffers little harm from many freezings.

Mineral Wedging. Most rocks in contact with water and the atmosphere are decomposed. This results in chemical additions to the materials composing the rocks. New minerals are produced of which the aggregate volume is greater than that of the old, and these then function as wedges to split adjacent minerals apart. The rifting may also be done by minerals forming in cavities from materials held in solution in liquids that permeate the rocks. Thus, a salt solution may form crystals of salt in pores of a rock and ultimately split the rock apart. This is done on a considerable scale in nature by crystallization of salt in rocks about salt lakes. The effects are like those due to ice. These processes are responsible for some exfoliation, as both freezing of water and formation of minerals take place in the outer parts of rocks and rift off fragments parallel to the surface.

Root Wedging. Many textbooks have illustrations showing the rifting effects of roots, but in most cases it seems probable that roots do very little work of this kind. Cracks must exist before roots can enter, and then all that roots may do is to widen existing cracks, and that seems possible only when the overlying weight is small.

Diastrophism. Diastrophism is responsible for great volumes of rock particles. These are produced along faults and where rocks have been sharply folded. Blackwelder has stated that in the arid regions of California and Nevada, diastrophism and the wedging effect of minerals are far more effective in producing rock particles than all other agents combined. Whether this statement has general application remains to be determined, but it is certain that many rocks are broken into fragments by crustal movement long before they reach the surface.

Undermining. Under many conditions, as in caves, on cliffs of the shores of lakes, riverbanks, and the sea, and on cliffs of arid regions, rocks are undermined by the action of water or wind. Collapse ultimately takes place with more or less rock breaking as a consequence.

Abrasion. Abrasion is defined as wear caused by rubbing. This is done in streams, on beaches, beneath glaciers, in deserts, and in the intestinal tracts of animals. The products have the fineness of flour, and much is of colloidal dimensions.

Experimental abrasion of feldspars was done by Daubrée, and impalpable mud was obtained. Abrasion of granites reduced feldspars to fine muds, micas to fine shreds, and rounded quartz grains to minimum dimensions of 0.25 mm. Rates of abrasion have been measured by Wentworth and Marshall (1920). In Wentworth's experiments, particles of limestone, of which each weighed 180 grams, were rolled 700 miles. The quantity abraded averaged 1 gram for each 4 miles. Marshall, through experiments with mixed gravels at a rate of travel of 1 m.p.h., found that hard rocks were abraded to the extent of a little more than 1.5 per cent by weight in 24 hr. In another experiment he used 5,000 grams of mixed gravels with 2 liters of water in a Deval machine. The gravels traveled at the rate of approximately 1 m.p.h. for 24 hr. There were produced 307 grams of materials with diameters less than 0.07 mm. It is obvious that the abrasion produced by flowing waters is of large magnitude.

Gregory has described abrasion of boulders in the beds of streams. This is greatest on the upstream side. The rate was determined by boring holes in the rocks to the depth of 1 in. These were materially shortened in a single season. Downstream sides of boulders were little affected.

Data do not permit statement respecting the agents most effective in abrasion, but it is suspected that ice abrasion has the greatest quantitative importance. At any rate, the products of ice abrasion are obvious in the immense quantities of rock flour discharged by melt waters from glaciers. However, there is no doubt that water currents are responsible for much abrasion, as one may see in the rounded boulders that are produced thereby. Much abrasion is also done by wind. The products of stream and wind abrasion are carried away gradually and are rarely observable in one place, as is the case at the end of a glacier where melt waters escape.

The particles produced by abrasion are extremely small, and many are of colloidal dimensions. These tend to float indefinitely.

Grinding. Grinding is a term applied to the crushing of particles by others that are larger. The products range from rock flour to larger dimensions, with the maximum dimension much below that of the particles to which the grinding is due. This work is probably more effective in streams and on beaches than elsewhere. Small rock particles associated with large on beaches with strong wave action have little or no possibility of survival, and Marshall (1920) has stated that fine gravel 2.5 mm. or less in diameter associated with other gravel with dimensions from 18 mm. or upward "cannot live" on a beach where waves keep gravel in movement.

Impact. Impact is the striking of rock on rock. Dimensions of the particles produced depend somewhat upon the dimensions of the impactee and the impactor. Particles of quartz are tiny flakes with sharp edges and curved surfaces. There is small volume and much surface. Places of contact have small crescentic fractures, known as *chattermarks.* Impact is mostly done in streams and on beaches. Marshall's studies showed that impact proceeds more rapidly than abrasion "when the impactor has ten times or more the diameter of the impactee," and that rocks may be broken by impact sixteen times more rapidly than by abrasion, and that grinding is 2½ times as rapid as impact. Under other conditions the ratios would probably be much different.

As consequences of the various methods of rock breaking or disintegration, particles are produced that from the point of view of dimensions may be placed in three classes: (1) particles ranging from colloid to silt size (Marshall, 1928), (2) particles ranging from sand to granule size, and (3) particles ranging from the dimension of pebbles to blocks of large size. If no other agents acted, these would be the sediments and they would contain the same minerals as the rocks broken, but concomitantly with the separation of particles from parents the processes of rock decomposition act upon them.

ROCK ROTTING (DECOMPOSITION)

General Considerations. Decomposition implies chemical destruction and production of substances different from the originals. It is an attempt to form minerals which are in equilibrium with environmental conditions. Some minerals, such as quartz and a few others, escape decay, but most common minerals of igneous and metamorphic rocks change to new products, the change ordinarily being attended by chemical additions of water, oxygen, and carbon dioxide. Most sedimentary rocks are composed largely of minerals produced by decomposition of other minerals, and further decomposition is impossible for these. However, many sedimentary rocks pass into solution in whole or in part and these rocks disappear or crumble.

Quartz does not readily decay and usually enters the products little changed. Feldspars change to carbonates, clay minerals, and silica. The ferromagnesian minerals change to carbonates, iron oxides, and silica, and some also produce clay minerals. Decomposition of the micas produces carbonates, clay minerals, silica, and iron oxides. Under certain conditions the clay minerals undergo further alteration to produce aluminum hydroxides and silica. The limestones and various evaporites pass into solution, leaving a residue of impurities consisting of silica, clay minerals, and iron hydroxides. If limestones contain chert, this also is in the residue in the form of large and small fragments. Sandstones are composed largely of quartz, and on removal of the cement they crumble into sands. Mudstones consist of fine particles of numerous minerals and crumble into muds with alteration of any substances that may have escaped an earlier decomposition.

The materials produced by decomposition are both heavier and of greater volume than the rocks from which they were derived. This is because something has been added in the decomposition, and the new substances are usually less compact than the original. The additions consist mostly of oxygen, carbon dioxide, carbon, water, and sulphur oxide. One hundred grams of the average igneous rock on complete decay has added 5.30 grams of carbon dioxide, 1.99 grams of water, 0.30 gram of sulphur trioxide, 0.72 gram of carbon, and about 1 gram of oxygen to make a total of 9.31 grams (Leith and Mead). Rarely may all the products be found at the places of decay. Usually the soluble and much of the colloidal materials are removed about as rapidly as formed, the residue remaining is oftentimes only a small fraction of the original rock. The losses include carbonates and sulphates, which are removed in solution, and oxides and hydrous silicates, which are removed mainly as colloids. That which remains consists mostly of insolubles and larger particles. Solution of limestones leaves behind only a part of the insoluble residue, and this may be only 1 or 2 per cent of the original rock. The residue from sandstones and conglomerates may be about equal to that of the original rock in volume, but there is likely to be a decrease in weight due to loss of cement.

Decomposition of igneous rocks results in much loss of silica, lime, magnesia, soda, and potash. There is usually considerable potash retained by the residue, which is assumed to be due to the formation of insoluble minerals, of which this oxide becomes a part or to adsorption of the potassium by some minerals of the residue. There is invariably a gain in water. It is assumed as a basis for comparison that alumina suffers no losses, but it is probable that such is not the case and that there are actual and often important losses of every constituent. The total loss is unknown.

There are different degrees of decomposition. *Mature decomposition* is considered to have been produced if only very resistant minerals of the original rock remain. This is probably rarely the case, but it may take place in warm humid regions covered with vegetation and of moderate relief. Under average conditions decomposition is not completed, and considerable parts of the original minerals may escape. If such minerals as feldspars, amphiboles, pyroxenes, olivine, and magnetite escape decomposition, it may be defined as *immature*. Immature decomposition is characteristic of lands with dry climates, excessive rainfalls, little vegetable cover, and moderate to greater relief. Immature decomposition is also found in sediments eroded from banks and beds of streams, from coasts of lakes and the sea, and in the materials produced by glaciers.

As sedimentary rocks ordinarily have passed through at least one cycle of previous decomposition, the imposition of a second cycle carries the decomposition to a greater extent toward maturity, so that there are fewer original minerals of igneous rocks in the products of decomposition derived from sedimentary rocks, and those minerals of the original igneous rocks that do remain are of the most stable character as related to the conditions. The extent, therefore, to which original minerals of igneous rocks are present in sediments affords some measure of determination of the immediate parent or parents of the sediments. If the sediments contain many original minerals of igneous rocks, it may reasonably be postulated that the provenance was an igneous terrane. If, on the other hand, the sediments contain only extremely resistant minerals of the original igneous rocks, it is probable that the provenance was a sedimentary terrane. If the sediments contain a suite of minerals that are characteristic of a single igneous terrane, it is reasonably certain that a single terrane is the provenance, but if the suite is one that is not possible in one variety of igneous rock, it should be postulated that two or more terranes constitute the provenance.

Agents and Methods of Decomposition. The agents of decomposition are the gases and moisture of the atmosphere, water and its contents, and organic matter. The work is mainly accomplished by the five processes of oxidation, carbonation, hydration, hydrolysis, and solution. There are a few other processes that take place, owing to the presence in some waters of sulphuric, hydrochloric, and other acids. Hydrochloric acid, chlorine, and sulphur gases are expelled by some volcanoes, and nitric acid is formed in the atmosphere by lightning. These substances are washed from the atmosphere by rain and produce some rock destruction, but of an undetermined quantity (Evans).

Living plants produce some rock decay. The roots of plants come in contact with rocks and act upon them in such a way as to extract something from them. This is probably done by carbon dioxide secreted by the roots, and the growing of plants in soils placed on a marble slab gives some slight measure of the effects. The ash of plants is a measure of the solids taken from the ground, and it is probable that some of these may be taken directly from rocks. Many rocks are covered with algae, lichens, and moss. Many lichens and algae are in small pits of 0.2 to 0.6 mm. in diameter and 0.1 to 0.2 mm. depth. These were certainly made by the plants which dwell in them, and it is probable that the material removed was utilized in whole or in part by the plants. It seems to have been demonstrated that some bacteria are connected with rock decomposition, and it is suspected that the role may be a large one. It has been proved (Thiel) that decomposition and solution of several kinds of rock are accelerated if they are placed in water containing soil bacteria. It was suggested that this acceleration may be due not to direct action of bacteria but to carbon dioxide generated by them. Studies of building stones of some of the public buildings of Great Britain and some of the quarries showed that the number of bacteria per gram of rock ranged from none to over 2,000,000 in a stone from the north wing of Buckingham Palace, London. The work indicated that the bacteria found produce carbon dioxide, some sulphuric acid, and some nitrogen compounds, all of which have decomposition effects upon the rocks inhabited by the bacteria (Paine, Linwood, Schimmer, and Thrapp).

Dead organic matter provides circulating waters with much carbon dixoide, and bacterial fermentation produces acids, among which are butyric, succinic, lactic, valeric, and the so-called humic and ulmic acids. Plants also produce various acids during life, among which are malic, citric, acetic, and others. All these compounds produce some decomposition and give increased solvent power to the waters in which they occur. Peat waters have been demonstrated to be effective solvents of iron, manganese, and silica, and waters high in organic matter readily transport these substances, supposedly because of the presence of organic colloids and acids and protective colloids that retard or prevent coagulation.

Certain kinds of decomposition promote others. Oxidation of such sulphides as pyrite, marcasite, galena, etc., results ultimately in the production of sulphuric acid, the presence of which increases decomposition of adjacent minerals. The solution of alkaline carbonates increases the solvent effect of water upon silica.

It has recently been learned that acid clay, a decomposition product of silicates, promotes a very rapid decomposition of some minerals. Studies made by Graham with powdered anorthite, hornblende, augite, microcline, biotite, and quartz in intimate contact with acid clay for 107 days showed that anorthite transferred 3.4 per cent of its calcium to the clay. Effects were less, but important with hornblende and augite. There was little effect on biotite and none on quartz. The studies indicated that colloidal acid clay is an effective weathering agent with respect to certain minerals and can remove about a hundred times as much calcium from anorthite in a given time as the hydrolytic action of ordinary water in equilibrium with carbon dioxide at atmospheric pressure.

Minerals and rocks composed of carbonates lose these by solution in water, particularly if the water contains carbon dioxide. Thus, rocks composed largely of carbonates become honeycombed by solution passages, and rocks of which the carbonates form the cementing materials crumble when the cements are removed. Rocks of which sulphates are constituents lose these by solution, and rocks formed of chlorides rarely appear at the surface, except in dry regions, because of solution.

Carbonation affects rocks in which uncarbonated sodium, calcium, magnesium, postassium, and ferrous oxide are present. Oxidation affects few minerals and rocks, as in most cases the combinations are satisfied with respect to oxygen, but minerals containing ferrous iron, such as the amphiboles, pyroxenes, biotite, olivine, magnetite, ilmenite, or minerals containing iron uncombined with oxygen, as pyrite, marcasite, and chalcopyrite, readily undergo oxidation.

Hydration, or hydrolysis, affects all substances containing silica, alumina, ferric iron, lime, magnesia, soda, potash, and other bases.

There are very few minerals that do not yield to one or more of these processes, and few rocks that do not contain some decomposable minerals.

Depth of Decomposition. Decomposition, in contrast to disintegration, can extend to considerable depths. The presence of a vegetable cover *retards disintegration*, whereas it *favors decomposition*, in that it compels more water and dissolved atmospheric gases to enter the ground, and the ground cover of dead organic matter through which the waters must pass provides them with an abundance of carbon dioxide and organic acids to facilitate the work of decomposition. Decay theoretically may proceed to great depths, but it is thought that most of it is done above the water table and for a slight distance below, owing to the inability of oxygen and carbon dioxide to penetrate far below the water table. Humid regions of little relief have the water table at or near the surface, and depth of decay seems to be quite shallow. Humid regions of much relief have a thick zone of decay, and semiarid and arid regions with very deep water table permit the atmosphere to penetrate to great depths and, with moisture condensed upon the surfaces of the rocks, to oxidize, carbonate, and hydrate all minerals subject to these processes. Such minerals rarely reach the surface of dry regions undecayed.

Great depth of decay is known in many parts of the world. Igneous rocks and the metamorphic equivalents of igneous and sedimentary rocks in the Piedmont Region of eastern United States south of the region of glaciation have exposures showing original structural and textural features but so decomposed that they may be removed by hand. It is recorded that decomposition in places has progressed to a depth of 24 meters. Decomposition in northwestern Georgia is said to have progressed to a depth of 60 meters, and in Brazil shales are reported to have decayed to a depth of 118 meters.

Decomposition of a rock begins on all exposed surfaces, of which one is in contact with the overlying soils, or atmosphere if no soil is present, and underground along such structural planes as joints. If a rock is permeable, decomposition may begin on the surfaces of all interstices, but as a rule the waters in these may move so slowly and exhaust carbon dioxide and oxygen so soon that decomposition is retarded to such an extent that it may be disregarded, and attention may be confined to decomposition on the surface and along structural planes. The weaker places and weaker minerals naturally yield first with the result that surfaces become irregular. As the rocks of the outer parts of the earth's crust are everywhere jointed to some degree, the result is that each block bounded by joints is decomposing from its periphery toward its center. Thus, each block is separated from that adjacent by a variable thickness of decayed and semidecayed rock. The core of the block becomes of a more or less spherical shape and may be sound or semidecayed, but the outside parts are formed of concentric bands that show increasing decay from the inside toward the outside. These more or less spherical bodies are termed *boulders of weathering* (Fig. 22).

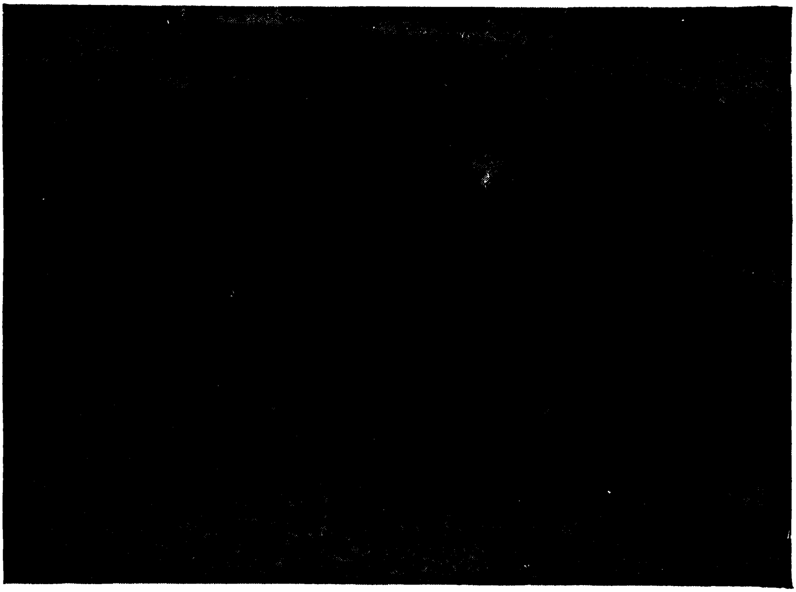

Fig. 22. Diorite showing spheroidal weathering and boulders of weathering. Road cut near Riverton, Sierra Nevada Mountains, California. (*Photograph by Eliot Blackwelder.*)

DISINTEGRATION VERSUS DECOMPOSITION

Considerable has been written respecting the relative importance of disintegration and decomposition in the production of sediments. Much of this has little value. In all regions where plants cover the surface the processes of disintegration—diastrophism and growth of minerals excepted—have little or no chance of operation, as the rocks are more or less decomposed long before they reach the surface. In cold regions where the temperature and high-water table retard and may prevent decomposition, disintegration effects are more obvious, and the same is true in high mountains where materials are quickly removed by falling, sliding, rain wash, or deflation. In warm regions, whether wet or dry, unless the dryness is of extraordinary rigor and the air is so dry that it does not condense moisture underground, decomposition ordinarily is so far in advance of disintegration that fresh rocks subject to decomposition rarely reach the surface. Finally, if a region is underlain by sedimentary rocks composed of sedimentary minerals or the resistant minerals of igneous rocks, there is nothing to decay, and, if there is not sufficient water for solution, the products mantling the surface are dominantly the result of rock breaking.

In general, under average climatic conditions disintegration is most extensive under conditions of much relief, decomposition under conditions of moderate relief. There is little disintegration and little decomposition very far beneath the ground-water table. Under most, if not all conditions, the thickness of the products of disintegration and decomposition beneath a base-leveled surface should be small.

Soils as Sources of Terrigenous Sediments

No consideration of the sources of terrigenous sediments is complete if soils are omitted, as in many, and perhaps in most cases, sediments are not derived directly from rocks, but from the products of decomposition, which as soils mantle the surface to variable thickness. Parent rocks supply sediments directly only on sea and lake shores, on riverbanks and beds, where glaciers erode, in extremely dry or very cold regions, and in regions where the slopes are so steep that the products of decomposition and disintegration cannot remain on the surface, so that it is bare and the rocks are exposed. In all other regions, detachment of particles from parent rocks is underneath coverings of particles previously detached, which in turn are under coverings of plants. The result is that the particles last detached remain where they are to undergo additional changes produced by the roots of plants, the burrowing of animals, and the decomposition effects of circulating waters and their contents. Finally the remains of the burrowing organisms and the roots of plants and the products of the organisms are mingled with the detached particles to form soil.

The minerals of soils are derived from parent rocks and to greater or less extent are products of decomposition formed from original minerals. Minerals directly derived from parent rocks usually are present to some degree. Quartz is by far the most abundant, and in many soils it is the dominant constituent.

Other original minerals are some of the feldspars and micas, zircon, tourmaline, and other rarer minerals. The quantities and varieties of minerals other than quartz and a few other minerals depend upon the maturity of decomposition. If decomposition has advanced to maturity, the quantity and kind of original minerals may be few. Common minerals of decomposition are various hydrous aluminum silicates, generally known as the clay minerals; iron hydroxides, usually termed limonite; less commonly iron oxide or hematite; aluminum hydroxide; carbonates; and still less commonly, sulphates. Carbonates and sulphates are not likely to be present in regions of considerable rainfall, as solution easily removes them. In addition there are various compounds of phosphorous, potassium, sodium, and nitrogen and others, all more or less rare as a rule. The common minerals of soils are the following:

Quartz.
Clay minerals: kaolinite, montmorillonite, beidellite, and others.
Limonite and hematite.
Apatite.
Augite.
Biotite.
Calcite.
Chlorite.
Collophane.
Dahllite.
Dolomite.
Epidote.
Gypsum.
Hornblende.
Illite.
Muscovite.
Olivine.
Orthoclase.
Plagioclase.
Serpentine.
Siderite.
Talc.
Zircon.
Zeolites: complex hydrated aluminum silicates of calcium, potassium, and sodium.

The first three of the list compose most soils. Other minerals are generally more or less rare.

As already noted, the character of a soil depends upon climate. Regions that have an excellent cover of plants and a cool climate have the top soils leached of iron; hence they have white to light colors and are known as *podsols.* Regions of warm climate with good underground drainage tend to have the clay minerals altered to silica and aluminum hydroxide. The same conditions favor removal of the silica, leaving a residue composed of iron and aluminum hydroxide with a range for either constituent of from nothing to 100 per cent. A soil of this type is known as *laterite.*

Many of the inorganic substances of soils are of colloidal dimensions. These give hardness to the soils when dry and plasticity when wet. They adsorb gases and organic and mineral substances and function in base exchange. Very frequently the colloidal matter is present as a coating on larger particles. It is said that colloidal matter makes up as much as 20 per cent of sandy loams, 30 per cent of loams, 50 per cent of clay loams, and 75 per cent of clays.

Soils contain more or less organic matter, which consists of plant tissue, parts of animals, pieces of charcoal, coal, substances resembling solid hydrocarbons, and derived organic matter which constitutes the so-called organic acids (humic and ulmic). Much of the organic matter is of colloidal dimensions and forms the so-called humus. This has high adsorptive properties for water, gases, and various elements such as calcium, magnesium, and potassium. The presence of humus adds to the ability of water to dissolve and transport such substances as silica, iron, and manganese.

TABLE 4. ORGANIC CONTENT OF SOILS OF THE UNITED STATES

	Sandy soils		Clay loams and loams	
	Soil, per cent	Subsoil, per cent	Soil, per cent	Subsoil, per cent
North central states	1.84	0.76	3.06	1.07
Northeastern states	1.66	0.66	3.73	1.35
South central states	1.16	0.55	1.80	0.65
Southeastern states	0.93	0.41	1.53	0.73
Semiarid states	0.99	0.62	2.64	1.11
Arid states	0.89	0.64	1.05	0.62

Physically, soils are composed of sands and larger particles, all greater than $1/16$ mm. in diameter; silts (between $1/256$ and $1/16$ mm.); and particles of clay size (less than $1/256$ mm. in diameter). Proportions vary with the soil. Soils of dry prairies and very wet regions contain the most organic matter, whereas well-drained areas in not too excessively humid regions have little organic matter except in the immediate vicinity of the surface. Sandy soils have less organic matter than clays and loams and top soils more than subsoils. There is also less in arid and semiarid regions than in humid. This is shown in Table 4.

Organic matter of soils is important in decomposition and transportation of sediments, as the organic colloids and compounds aid in solution and serve as stabilizers in colloidal transportation, so that such substances as silica and iron, aluminum, and manganese compounds transported as colloids are not precipitated on meeting electrolytes or colloids of opposite sign. As soils yield their burdens to aqueous transportation largely from the top, it is almost inevitable that some organic substances become associated with the inorganic materials. This organic matter, when deposited with the inorganic sediments, brings about reduction of any ferric oxides that may be present.

SUMMARY

Most sediments are of terrigenous origin, and even the bulk of the organic sediments is ultimately referable to terrigenous sources for the various substances of which they are composed. Furthermore, the volcanic and cosmic materials after decomposition produce sediments that are not different from those directly referred to a terrigenous origin. Hence, a consideration of terrigenous sediments gives results that are applicable to all other divisions with the exception of the postulated magmatic sediments of which too little is known to validate any statements.

The terrigenous sediments are divisible into two divisions: particles derived from original rocks and decomposition derivatives from original rocks. The former may be of any composition, as the parent rocks may be igneous or sedimentary or the metamorphic equivalents of these, and hence any mineral of any of these rocks may belong in the first division. Minerals of the second division are mainly carbonates, sulphates, chlorides, oxides and hydroxides, and hydrous aluminum silicates, generally known as clay minerals. Long and mature decomposition generally reduces any rock to carbonates, oxides and hydroxides, and hydrous aluminum silicates and a few rare minerals of the original rock that do not decompose, so that the mineral mixture of a rock in process of decomposition becomes more simple the nearer the products approach mature decomposition.

The mature decomposition of an average igneous rock has been estimated as yielding products that would make sedimentary rocks to the extent of 82 per cent shales, 12 per cent sandstones, and 6 per cent limestones (Leith and Mead). Mature decomposition of the average igneous rock involves addition of material from both atmosphere and hydrosphere, so that 100 grams of the average igneous rock yields 107.4 grams of sediments, which are composed of 87.8 grams of average claystone and siltstone, 12.9 grams of average sandstone, 6.7 grams of average limestone, and, in addition, 6.6 grams of sea salts or a total of 114 grams. The volume is increased from 100 units to 112.4, exclusive of pore space (Leith and Mead). The changes in mineral content are shown in Table 5.

TABLE 5. MINERALOGICAL COMPOSITION OF COMBINED SEDIMENTS AND AVERAGE IGNEOUS ROCK

Mineral	Average igneous rock, per cent	Combined sediments, per cent
Quartz	20.40	34.80
Albite	25.60	4.55
Anorthite	9.80	
Orthoclase	14.85	11.02
Biotite	3.86	
Muscovite	3.85	
Hornblende	1.60	
Augite	12.90	
Olivine	2.65	
Magnetite	3.15	0.07
Titanite and ilmenite	1.45	0.02
Kaolin	......	9.22
White mica	......	15.11
Chlorite	......	5.29
Limonite	......	4.00
Dolomite	......	9.07
Calcite	......	4.25
Gypsum	......	0.97
Rutile	......	0.55
Apatite	......	0.35
Carbon	......	0.73
Total	100.11	100.00

This table serves to show the great contrast in mineral contents of igneous and sedimentary rocks and that only a few igneous minerals survive decomposition. The white mica probably contains some allogenic muscovite, as this mineral is frequently known to escape decomposition. Kaolin should be considered to represent all the clay minerals. The ferromagnesian minerals and most of the feldspars disappear after fairly mature decomposition. Micas resist decomposition fairly well, particularly muscovite. Magnetite and ilmenite yield readily in warm, humid climates, but in the cooler climates of the polar parts of the Temperate Zones these and other easily decomposable minerals are frequently seen in sediments. Although apatite is not indicated in the average igneous rock, it is present and generally yields readily, but it may enter sediments without much alteration. Ilmenite and titanite frequently alter to rutile, anatase, brookite, and leucoxene. Zircon, tourmaline, and a few others not shown in the average igneous rock undergo little or no change.

TABLE 6. REDISTRIBUTION OF THE OXIDES OF THE AVERAGE IGNEOUS ROCK*

Oxide	Mudstones, per cent	Sandstones, per cent	Limestones, per cent
Silica (SiO_2)	83.0	16.5	0.5
Alumina ($Al_2O)_3$	95.3	4.3	0.4
Iron (Fe_2O_3, FeO)	96.5	2.9	0.6
Magnesia (MgO)	76.0	5.3	18.7
Lime (CaO)	43.6	11.2	45.2
Soda (Na_2O)	95.0	4.7	0.3
Potassium (K_2O)	93.8	0.7	5.5

* Leith and Mead.

The redistribution of the oxides of average igneous rock in the three classes of sediments is shown in Table 6.

The table shows that, except for lime, each oxide has a greater distribution in the mudstones than in the sandstones and limestones and that there is more lime in the combined sandstones and mudstones than in the limestones. The analyses on which the figures of distribution are based were made of rocks which must necessarily have been derived from no great distance beneath the surface for most of them. If all samples analyzed could have been derived from deep beneath the surface, it is possible that somewhat different results might have been obtained. Further, no consideration could be given to probable vast thicknesses of sediments beneath the deep sea, and these might show a somewhat different distribution. The figures are interesting, but not too much weight should be placed upon them.

It has been stated that the magnesia in the sediments fails to account fully for the quantity in the average igneous rock and that there is an excess of lime. These discrepancies need cause no concern, and if there is actually a deficiency of magnesia and an excess of lime in the sediments exposed on the continents, there are no reasons for not believing that deep-sea sediments may contain an excess of magnesia, and there are good reasons for believing that Paleozoic and early Mesozoic deep-sea sediments may have a deficiency of lime.

Sedimentary rocks as sources of sediments yield essentially the same things as those of which they are composed except to the extent that these rocks contain minerals from the original igneous rocks. These minerals continue along the road of further decomposition. Many limestones, particularly dolomitic limestones, contain much chert in the form of small and large particles of authigenic origin. This breaks up on solution of the containing rocks, and as the cherts not uncommonly contain considerable quantities of included carbonates, these are dissolved, if accessible, and the chert is rendered more porous. Chert and flint, as later stated, are composed of a mosaic of quartz and chalcedony, of which the latter is the more soluble. There may also be some opal, which is more readily removed than the chalcedony. Removal of any of these constituents ultimately leads to crumbling and the formation of a powder which is red to white depending upon the quantity of iron oxide present. This may be a decomposition product from siderite or pyrite in the chert, or it may be introduced from without.

Decomposition of sandstones produces sands like the parent rocks, and shales decompose into muds.

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CHAPTER V

INTERRELATIONS OF ORGANISMS AND SEDIMENTS

The interrelations of organisms and sediments naturally fall into the three divisions of (a) *influence of sediments and sedimentary conditions on organisms,* (b) *modification of sediments by organisms,* and (c) *production of sediments by organisms.* These three divisions are interrelated in that the organisms in many cases make the sediments on which they live; they modify the sediments they produce as well as all sediments of other origins; and the organisms may be permitted to remain or may be expelled by the sediments and the sedimentary conditions they have produced. Expulsion prevents further production or modification of sediments at the place by these organisms, but others may enter and produce other sediments and perhaps other modifications.

INFLUENCE OF SEDIMENTS AND SEDIMENTARY CONDITIONS ON ORGANISMS

Environmental conditions determine whether organisms live therein or are killed if they enter. Environmental conditions may be optimum for some organisms, tolerable for others, and deadly to a third group. Thus, certain animals live in fresh water; fresh-water conditions are not liked but are tolerated by a second group; and death is a consequence if typically marine organisms enter.

Any organism in the course of its life and duration of its remains afterward is concerned with four environmental situations, which are: (a) *environment of life,* (b) *environment of death,* (c) *environment or environments through which the remains may be transported after death,* and (d) *environment of entombment.* Many organisms live, die, and are entombed under the same environmental conditions, but others die far from the places where they normally lived, and they may be entombed neither where they lived nor died. These possibilities should be realized.

Interpretations of the significance of organisms found in rocks hinge in the first place on whether they lived in the places where they are found or whether they lived somewhere else. That is, do the rocks represent the places where they died, or did they die elsewhere and is this merely the place where they were buried? In some cases these different things may be determined, but in most cases existing knowledge does not permit this to be done. Students too often have postulated that the places where fossil organisms are found are the places of life, death, and entombment.

The history of any organism may be placed in the two phases of life and death, and reconstruction of the life relationships of fossil organisms cannot be made if only the places of death are known, and results may be extremely erroneous if the places of entombment are assumed to be the places of life. The life relationships of organisms are the *biocoenosis;* those relationships that begin with death and terminate with entombment are the *thanatocoenosis* (Wasmund). During the time of the biocoenosis an organism to a greater or less extent may control its actions; after death no such control exists, and the body or its parts are under the control of inorganic and organic factors external to the organism. These may transport the remains through and to environments where the organism would not or could not have gone in life. This may be illustrated by considering the cases of marine organisms dwelling in waters of a certain depth. Death takes place; the remains may be transported to deeper or shallower waters; they may be picked up by waves and thrown upon a beach; and strong storm waves may sweep them over littoral barriers and leave them high and dry on land. If the remains have food value, birds or other animals may carry them inland and leave shells long distances from water. Live shells may be similarly picked up and thrown on shore or carried elsewhere into environments that compel death to take place. That this is not an uncommon event may be readily proved by a visit to any seashore, where the finding of shells of marine animals several miles from the sea emphasizes the significance of the thanatocoenosis. Or different environmental conditions may be brought to an animal's home; an example is the great rain that fell on the Great Barrier Reef of Australia in 1918, nearly 90 cm. (35.7 in.) in 8 days, which so freshened the waters as to kill all animals in places to a depth of 3 meters (Yonge).

Marine planktonic animals frequently drift into estuaries and marginal lagoons. Many of these contain brackish, fresh, or excessively salty waters. Death may result. Pools of the tidal zone may be filled with animals at each high tide. These pools have waters of normal saltiness when the tide is high but may become filled with fresh water as the tide falls; death results for most, if not all, of the unfortunate animals. It cannot be doubted that millions of animals are killed in this way annually in places where they did not live. The Lower Paleozoic graptolites are preserved in greatest abundance in black shales—a product of reducing conditions and foul waters. It is very doubtful if the graptolites normally lived over these bottoms. Certainly they could not have lived on them. The graptolites were in considerable part plankton and epiplankton, and as such they dwelt in greater or less abundance over all bottoms. They were killed in largest numbers under abnormal conditions of sea water. Such conditions do not occur frequently in the open ocean but are common in shallow waters where replenishment of oxygen is difficult and where changes of salinity are frequent. The souplike, sapropelic bottom materials afforded excellent places for entombment, as they were foul because of lowness of oxygen and were thus without scavengers. Deposits of greatest abundance of graptolites should not be termed those of "graptolite seas" but those of graptolite killing.

Organisms dwell where they do because the environmental conditions permit them to do so, and they have living positions adapted to the places where they dwell and to their life habits. These environmental conditions are factors in sedimentation, and hence organisms and their positions and everything else that may be learned about them may serve as aids in interpretation of the significance of the containing sediments. Much organic matter ultimately becomes sediments, and the positions and characters of this organic matter are important in reconstruction of ancient environments. Important environmental factors of aquatic organic control and also control of the deposition of sediments and their characters are the substratum, chemical characters of the waters, turbidity, food supply, depth, temperature, and competition of other organisms.

The substratum varies from place to place and from time to time at the same place, and with these changes go changes in the benthonic organisms and to some extent in the plankton and nekton of the overlying waters. The substratum is an environmental result, and the organisms that dwell upon and within it are affected by previous environmental factors. Lateral variations in the substratum lead to discontinuous lateral distribution of organisms, and variations in the substratum in the time relation lead to vertical discontinuous distribution of organisms to the same degree.

The materials of an aquatic substratum vary through a wide range. Some bottoms are composed of solid rock; others are composed of stationary or shifting sands, heaps of shells, or muds that have settled compactly; and others are composed of muds that are in the form of a thick sludge or soup. Bottoms undergoing severe erosion have benthonic organisms in protected areas or spots. Solid bottoms with sufficient wave and current wash to maintain an adequate supply of oxygen but not enough seriously to erode the bottom support large populations. A solid bottom may be composed of rocks, sands, shells, or solid muds. The need for solid bottom is shown by the practice of oyster fisherman—the spreading of empty oyster shells over the sea bottom to provide a solid base to which young oysters may attach themselves. Bottoms composed of shifting sands support few macroscopic organisms. The dangers of burial are great, and moving sands hamper the animals' activities. These places are often real aqueous deserts and are almost as difficult for organisms as the drifting sands of land deserts. Bottoms of this kind are not uncommon in the present neritic life zone, and perhaps some of the cross-laminated, poorly fossiliferous sandstones of the geologic column represent such bottoms. Bottoms composed of soft mud may have a small population, as such bottoms are places of active deposition, possibilities of burial are great, and there is always the danger of smothering caused by sinking into the muds. Strangely enough these deposits may contain many shells, as those that are present are easily buried and preserved from destruction by scavengers. Bottoms composed of sludge or souplike muds are perhaps the most difficult on or in which to dwell. Such bottoms are likely to be deficient in oxygen; in some cases the bottom materials and overlying waters are filled with hydrogen sulphide. Burrowing forms and anaerobic bacteria are the only benthonic organisms that may be present, but shells of plankton and skeletal matter of nekton derived from upper waters may be common.

Herdman states that in the Irish Sea the nature of the bottom is more important than depth, and "a muddy bottom will support a similar fauna at ten fathoms in one place and at fifty fathoms in another." The controlling influence of the character of the bottom on the kinds of animals dwelling on it is splendidly shown by Tauben Bank and surrounding bottom in the Bay of Naples. This bank rises to 50 meters below sea level and is surrounded by waters ranging in depth from 200 to 500 meters. The deposits on the bank in 1885, when it was studied by Walther, were calcareous sands upon which lived 360 species of animals, of which 310 species might occur in fossil form. The deeper waters surrounding the bank had mud bottoms on which lived 142 species of animals, of which only 45 species might be preserved as fossils. Only 14 species possible of occurrence in the geologic column were common to the two environments, and yet the horizontal distance between the two bottoms is small. Twenty-five years later the bank was again examined by Walther. In the meantime, the bottom had become covered with lime-secreting algae and the benthonic fauna had changed, but there had been no climatic or diastrophic event to which the change might be referred.

Changes in the character of sediments thus force changes in the bottom fauna, compel expulsion of those organisms adapted to the previous bottom, and permit entrance of a new association. The result is vertical discontinuity of organic associations, thus accounting for the variations of fossil content in successive strata of the geologic column, which in their lithologies prove deposition under different environmental conditions. Changes in character of bottoms are normal and recurring events in their history, and with physical changes faunal ones must follow. The changes in fauna from layer to layer seen in the geologic column should not be regarded as extraordinary, but normal. *The really extraordinary feature is vertical continuity in character of sediments and faunas.*

Bearing on modification of organic distribution and also on organic dimension are the chemical characters and gaseous contents of waters. Waters of low or high salinity or waters with high content of hydrogen sulphide do not permit the entrance of numerous forms of aquatic animals, and those that do enter and live may not be so large as individuals of the same species in normal marine waters. Thus, on the east shores of the Baltic Sea the common shells of *Mya arenaria* are not more than one-half to one-third so large as on shores of the open ocean. Aquatic species of plants acquire carbon dioxide directly from the waters in which they live, and growth varies directly with the carbon dioxide available, which in turn varies with the carbon dioxide tension of the atmosphere and the temperature, as more carbon dioxide is held in cold waters than in warm. The result is a greater growth of plant tissue in cold than in warm seas. The absence of oxygen in waters filling deep holes of the sea bottoms and in other places of poor circulation renders it impossible for these places to support a normal benthonic fauna of aerobic organisms, but anaerobic organisms, chiefly bacteria, may be abundant. Among the latter are those that reduce sulphates and form hydrogen sulphide and sulphuric acid. The waters thus become poisonous, and even shells that drop in from overlying waters may pass into solution. Nitrates in solution encourage growth of plants and invite the entrance of nitrate-reducing bacteria and the formation of ammonia.

Turbidity limits the presence of animals and plants because of mud settling upon delicate structures. This prevents acquirement of the oxygen needed for respiration of animals and more or less closes the avenues in plants by which they acquire carbon dioxide. Most organisms can endure a limited degree of turbidity, but if this passes limits, which are probably different for each species of organism, they are smothered. Where the waters are highly turbid, plant life is limited in depth because of difficulty of penetration of light.

Food supply is naturally among the greatest factors, if it is not the greatest factor, controlling the presence and migration of organisms. Animals are usually adapted in numbers and kinds to the food supply. They dwell in abundance on bottoms and in overlying waters where food is abundant, and are scarce where food is scarce. Plants and microscopic organisms are the basic food for aquatic animals, and where these are present in abundance there also are the animals with their parasites, commensals, and predators. The number of herbivorous animals is adjusted to the number of plants, and the predators are adjusted to their prey. Scavengers form the end of the chain of food, and they range from microscopic dimensions to animals of large size. These are also adjusted to the food supply and, if the dead matter is abundant, they are likewise. They may pass all organic matter and everything containing organic matter through their digestive tracts and may eliminate every evidence of organic origin in sediments. Nekton migrate with their prey just as the cod follows the squid and the whale follows the caplin and other small fish.

Depth is a factor in controlling distribution of organisms and does so on two bases. Penetration of light is inversely proportional to depth, and this limits plant growth downward, which in turn limits the downward distribution of those animals which directly or indirectly depend on particular plants. Each herbivorous form dwells at the level of the plants on which it feeds, and predators live at the depth levels of their prey. The organisms are also adapted to the pressures due to depth. Thus, the plants of the bottom are arranged on contours from the surface downward, each confined to certain depth limits. The various herbivorous animals are adapted to the same contours both by limits of depth and food supply and so to the last organisms dependent upon particular plants for food either directly, as herbivores, or indirectly, as predators. From its limiting depths an animal may venture only at its peril. These depths are wider apart for some animals than for others.

Temperature, in general, lowers from the surface downward except in the polar seas where the top waters may be the coldest, but both top and bottom waters are cold in those seas. Each variety of organism is adapted to a certain range of temperature, and this range may be unlike that of any other form. Organisms are adapted to all temperatures found in oceanic or other waters, even to those in oceanic waters which may be below freezing of fresh waters, as there are inhabited waters on the bottom of the North Sea, and probably elsewhere, that are below 32°F. Contrary to what might be expected, the quantity of life in cold waters is not less than in warm, and "animal life of all kinds, both in the surface waters and on the bottom," is more dense in polar waters than it is in tropical ones (Allen). The variety of life seems to be less. The abundance is probably referable to the greater quantity of carbon dioxide in cold waters than in warm and a more abundant plant growth due to this fact. Animals that venture or are carried into waters beyond the limits of their temperature range rarely return, and if the water over a bottom changes temperature rapidly, as happens when a storm sweeps cold waters so as to displace warm, the destruction of life may be cataclysmic (Verrill). Some bottoms are almost barren of life because of rapid and frequent fluctuations of temperatures of overlying waters.

That animals of the present are adapted to a certain degree and range of temperature is no good reason for assuming that their ancestors or close relatives of the past had the same temperature limits. Most reef corals of today do not readily endure a temperature below about 68°F., and it has been frequently assumed that such was also true of the Paleozoic reef corals. The basis for any such assumption is extremely poor, and it should not be forgotten that there are many close relatives of the forms requiring a 68°F. temperature which live in waters that are very cold. The Paleozoic reef corals are not even closely related to modern forms.

Organisms compete with other organisms for space, food, etc., and organic associations present competition to other organic associations. Conditions may be extremely favorable for certain organisms, but they can not establish themselves because of the competition of other organisms or groups of organisms.

The impact of this range of sedimentary conditions and products is obviously complex, and this complexity may make it difficult to determine why particular organisms are at any particular place. In some places the association of conditions is such as to provide the optimum for the support of an organic association, and, where this is the case, the bottoms teem with benthonic life and its commensals, parasites, predators, and scavengers. In other places the combination of conditions may produce the very worst organic environment, and the bottoms are almost barren of life. Intermediate combinations of conditions produce environments that may favor certain forms, and these are likely to be present in abundance. The result is that shallow bottoms have few places where the impacts of the various sedimentary factors are the same, and thus the same association of organisms is not often duplicated. Wherever organisms dwell, however, whether under cold or warm conditions, the associations tend to be temporarily balanced and each reflects the environmental conditions of the place. Organic remains that do not show a community of ecologic character, and are thus unbalanced, represent either a thanatocoenosis, or a mixture of biocoenotic and thanatocoenotic remains.

Thus, there are coral-reef faunas, as over the Great Barrier and other reefs of the present and over the Silurian reefs of the Michigan Basin. Colonies of *Pecten* are present in the shallow waters of parts of some coasts and are absent elsewhere, just as local banks of *Myalina* are present in the Pennsylvanian of Kansas. There are hordes of *Mytilus* on a part of a coast and they are wanting on other parts, just as in the Silurian of Anticosti a given horizon in part of its distribution is filled with *Virgiana* and the genus is rare or absent in other parts of the distribution of the formation.

All groups of organisms are related to sedimentary factors and to sedimentary results consequent to these factors. These relations are discussed on the pages that follow. Certain groups are not given a great deal of attention, as they add little or nothing to the formation of sediments and leave little or no evidence of their onetime presence. Hence, they do not aid in reconstruction of sedimentary environments. Little is said of some groups for the reason that little is known respecting the impress of the sedimentary conditions upon them. This is regrettable, but the ecologic relations have not been studied, and until that has been done little can be said that has value (Allan, Allen, MacGinitie, Sumner, Twenhofel).

PLANTS (ARNOLD, BERRY)

Both land and aquatic plants live in societies or associations. Each society is complex, for the plants in the association are adapted not only to the physical and chemical factors of the environment but also to each other. These associations pass laterally into others through transition associations, and each association is succeeded in time by another. The succession may be abrupt or transitional.

Before fossil plants may be used to reconstruct a physical environment it must first be determined whether the plant material is *autochthonous*, that is, whether it grew where it is found; or *allochthonous*, that is, whether it was transported to where found. Plants float readily and for long times, and thus may be readily transported miles from places of growth and be deposited in river, lake, swamp, lagoonal, estuarine, or marine deposits. The transported plant materials may bear absolutely no ecological relation to the region of deposition other than the chance incident of lodgment or sinking. The finding of land plants on the bottom of the deep sea many miles from land illustrates the extent to which erroneous conclusions may be drawn if the autochthonic or allochthonic character of the material is not first determined.

Plant matter buried at the places of production bears evidence that may aid in reconstruction of the conditions of growth. Autochthonic plant materials tend to be little worn, to exhibit perfect leaves and other structures, and to have fruit and foliage in association. Stems and trunks may be in the positions in which they grew, roots may be in place, and the several varieties of plants should show a community of ecologic characters, whereas transported plants would not have stems and roots in place, fruit and foliage would not be likely to be associated, leaves would be frayed and stems and trunks would be abraded, and the association would not be likely to be a natural one. Desert, mountain, and swamp vegetation may be present in the same stratum. The longer the transportation, the greater would be the wear and mutilation,

Aquatic plants have very little value in reconstruction of ancient sedimentary environments, as they are usually so poorly preserved as not to be identifiable. If autochthonic, they may indicate whether the waters of deposition were fresh or salty, and they show that the sediments in which they are contained were deposited in depths within the influence of light. The chief fossil remains of aquatic plants are the calcareous structures produced by them through extraction of carbon dioxide from water. This forces precipitation of carbonates about the bases of the plants and upon their surfaces. It is impossible to tell from these structures whether the plants lived in fresh or salt water, as the same forms of structure are produced in both environments. The structures show little or nothing of plant structure; they merely tell of plant activity. It seems probable that the plants forming these structures attain greatest development in warm waters, but they are locally very abundant in the North Atlantic on the American side.

Bacteria, diatoms, and fungi dwell nearly everywhere in fresh and salt waters and in soils in species and abundance adapted to the environmental conditions, but knowledge has not advanced sufficiently to permit reconstruction of ancient environments on the basis of the presence of any of them. Diatoms secrete siliceous tests which are abundant on parts of the deep bottoms of the ocean, particularly in the colder waters of the southern hemisphere, and on the bottoms of many lakes.

The higher forms of terrestrial plants are far more useful in reconstruction of sedimentary environments than are the bryophytes and thallophytes. On the lands of the present, and presumably on those of the past, plants are adapted to a wide variety of conditions of which many certainly are reflected to some degree in the sediments. In many cases the adaptations are shown in roots, stems, and leaves. Leaves possess specialized stomata, branches have spines, and the roots are tufted or long and vertical. Leaves of desert plants are small and have curled margins, waxy surfaces, palisade tissue, and small stomata, and many plants have horizontal root systems. Swamp plants have horizontal root systems and never have long tap roots. Some have knees and enlarged boles, as in *Taxodium* (cypress), and the leaves have abundant and delicate stomata. Some plants of moderately moist regions have widely spreading horizontal root systems, as in the beeches; others have roots like the oak that descend at various angles into the ground; and others, as the hickory, have long tap roots. Each variety has its temperature range; each is restricted to an alkaline, neutral, or acid soil; and each has its requirement of rainfall and atmospheric humidity. The ancestors of living plants are extinct, and there is little justification for, and no merit in, assuming that these ancestors were limited by the same restrictions as their living relatives. They may have been, but more definite statement is not permitted. Very close relationships permit postulates, but nothing more.

Small bodies of water with low shores, such as lakes and lagoons, may have poor circulation and are not likely to receive much vegetable matter that has traveled far. Plant remains buried in the deposits of these small bodies of water have a community of ecological relationships and may be excellently preserved with perfect and unabraded leaves. Similar small bodies of water surrounded by hilly or mountainous shores may readily receive leaves, stems, and fruits from different elevations from water level to the hills or mountains. The association of plant remains is not likely to show a community of ecological relationships, but the fossil plants may be excellently preserved. Some layers may teem with plant remains, particularly if the region contained active volcanoes at the time of deposition, as showers of ash may fall upon the surface of the lake and carry floating leaves to the bottoms and at the same time bury them. An event of this kind took place during the Tertiary in the Florissant district of Colorado, where ash layers in ancient lake deposits contain thousands of leaves, other plant remains, and many insects.

If sediments contain plant remains of which many show wear in the form of frayed leaves, leaves without stems and petioles, and abraded stems associated with remains in excellent preservation, it may be postulated that the deposits are those of a delta, a large lake, or the sea, as may be indicated by other organisms in the deposits or by the sedimentary associations.

Plant remains deposited in water after some transportation have leaves expanded and parallel to bedding, scales of cones in contact, and nuts with hard coats closed and intact. Plant materials deposited by winds have leaves more or less curled and heaped together, scales of cones not in contact, and hard coats of nuts cracked and spread. Leaves curled at the times of deposition may be mashed upon themselves if sediments did not fill the curls.

PROTOZOA (CUSHMAN)

Protozoa live everywhere, but few have any possibility of leaving a fossil record. Only radiolarians, silicoflagellates, and foraminifera have preservable hard parts, so that inquiry into the influence of sedimentary conditions upon Protozoa may be limited to these three divisions.

The Radiolaria are and always have been exclusively marine, and most are plankton. They dwell in all marine waters, and their shells start to all bottoms in numbers determined by the luxuriance of their growth in the overlying waters. On bottoms to depths of 3,600 to 4,500 meters the tests ordinarily make little impression because of abundance of other sediments. Bottoms at depths greater than 7,500 meters rarely receive the shells because of their having passed into solution. Thus, between 3,600 to 4,500 meters and 6,000 to 7,500 meters radiolarian tests may make the greater part of a deposit. This fact has repeatedly suggested that radiolarian tests in significant quantities indicate bottoms within the depth range of around 3,600 to 7,500 meters. However, wherever conditions are such that other sediments are not making a significant deposit, it is possible for Radiolaria to do so, and these deposits may be made in extremely shallow water, as shown by radiolarian shale of the Pico formation of the Los Angeles Basin of California, which bears every evidence of deltaic deposition, and similar shallow-water radiolarian deposits on Trinidad. It is even possible for radiolarian tests to make shore deposits, and under a certain combination of favorable conditions the tests may be swept from the shore by wind and deposited on land. The mere presence of abundant radiolarian tests is not positive evidence of any depth of water and certainly does not prove deep water. Abundant tests merely suggest a deep-water origin, as the most important deposits are made there. The Radiolaria are now most abundant in warm waters, and this may always have been the case, but proof is lacking.

Beyond that the Silicoflagellata are marine and planktonic, too little is known of existing relations of the group to sedimentary conditions to permit venture of opinion respecting past relations.

Most foraminifera are marine, and all with hard tests live in that environment. They belong either to the plankton or benthos. The planktonic forms live in the sunlit upper waters, and the tests rain down upon all deep and shallow bottoms of which the depths are not so great that the tests pass into solution before that goal is attained. The shells are delicate and fragile, with much surface compared to volume, so that by the time they have sunk 3,600 to 4,500 meters there is little left. Most have passed into solution. In shallower depths the planktonic shells accumulate on all bottoms, but, if other sediments are abundant, foraminiferal shells form but a minor component of any deposit. At places and in depths where other sediments are in small quantity the foraminiferal shells have a chance to make a large part of any deposit. These are known as oozes, and they have their most extensive distribution in deep waters to 4,500 meters or less, with most of the deposits considerable distances from land. However, a foraminiferal ooze does not require deep water or water far from land for its formation, as the planktonic foraminifera live over shallow bottoms close to the shore in as great, if not greater numbers than over deep bottoms far from land; and if for any reason other sediments are not accumulating in large quantity to mask the foraminiferal tests, significant deposits may be made in shallow waters; and if currents are suitably directed to carry the tests to the shoreward part of the neritic zone and even upon the beach, significant deposits may be formed in those environments, as is the case on the coast of Trinidad and on the coast of Dog Bay in Ireland, where beach sands are made of foraminiferal tests. At that latter place the tests are rolled up by winds to form dunes that have traveled inland for several miles. These obviously do not indicate deep water.

Benthonic forms have large shells for the group and are most abundant in shallow tropical waters to depths of about 30 fathoms and with temperatures above 75°F (23°C.).

The work of Norton, Natland, and Lowman and his associates (data of Lowman mostly not published) show that different species of the benthonic foraminifera are related in distribution to depth and temperature. Glaessner has placed the depth zones in five groups, as follows:

Brackish water of lagoons
14 to 125 ft., temperature 21.4 to 13.2°C.
125 to 900 ft., temperature 13.2 to 18.5°C.
900 to 6,000 ft., temperature 8.5 to 4.0°C.
6,000 to 8,340 ft., temperature 4.0 to 2.4°C.

Species of the same genus live in different temperature and depth zones, and some of the species are stated to be closely related. Phleger has made a somewhat different temperature and depth classification. Whether benthonic foraminifera required the same conditions in the past is not known, but it may be assumed that fossil species closely related to living forms probably had similar conditions of habitat. Large forms are confined to tropical and warm waters; polar forms are small. This may also have been the case in the past. Foraminifera in highly alkaline waters acquire heavy shells, whereas if they live in acid waters, the same species form shells that are thin and chitinous. This seems a characteristic that is probably applicable to foraminifera of the past as well as to those of the present.

COELENTERATA (SAVILLE-KENT, VAUGHAN, YONGE)

Nearly all coelenterates are marine, and there are no reasons for believing that such was not always the case. Fresh-water forms are few in number, are relatively uncommon, and the remains are not likely to become fossilized. Most coelenterates belong to the benthos; a few are plankton. Existing coelenterates do not like to live in turbid waters or on muddy bottoms.

For purposes under consideration, coelenterates may be divided into the graptolites and the stony corals. The stony hydroids are included with the latter.

Graptolites are extinct. Some were plankton, others benthos. Planktonic forms seem to have been the more common. Some were attached to floating objects and were epiplanktonic. Planktonic forms seem to have lived in all normal marine waters and to have had wide distribution. Their dead remains must have sunk to all bottoms that were not so deep that the remains dissolved before arrival. Those that attained well-aerated bottoms must have been destroyed very quickly by scavengers, thus accounting for few graptolites in the deposits of such bottoms. Only if the dead graptolites settled on bottoms composed of materials not favoring the presence of much scavenger population would the remains have much likelihood of becoming entombed. Fossil graptolites are most numerous in black shales—the deposits of reducing conditions, of foul and poisoned waters, and of waters with few or no macroscopic scavengers. On such bottoms the chances for escaping destruction could not have been better. It is very doubtful if the waters over the black mud bottoms on which the graptolites collected at times in abundance were those in which the graptolites normally lived. It seems likely that these waters were not normally marine but of such character as to kill the graptolites when they entered, as might have happened when storm waters floated them into coastal waters of poor circulation, just as similar killing of marine animals is caused by present-day storms and takes place in a small way on many beaches with each rise and fall of tide. An extremely large number would be killed in a short time at more or less long intervals, thus accounting for the fact that graptolites are not generally distributed in black shales, but are very abundant in certain thin layers. Attempts have been made by some American and European students to refer the black graptolitic shales to deep-sea origin. Such attempts were made more often in the past than recent years. The writer sees little support for such a view.

The hydroid stony corals seem largely to belong to shallow waters, but beyond that general statement little more may be said.

Much more is known of the anthozoan corals. These are known to live in both deep and shallow waters and in both warm and cold waters, with species adapted to almost every normal marine condition. They prefer clear waters and bottoms with solid objects to which the young may attach themselves. The waters must be of normal and nearly constant salinity, as the corals are sensitive to changes. Whether they were always so is not known. Bottoms of muds and moving sands support few corals, and growth is greatly restricted and may be prevented if waters are muddy. Each species of coral has its particular depth and temperature range. Corals of deep water are small and have delicate exoskeletal structures, and the same species in the deeper parts of their ranges have more delicate structures than in the shallower parts.

Reef corals are confined to the neritic life zone and to depths which for most range to about 25 fathoms. About the Hawaiian Islands, 77 species of corals are in the range from 0 to 25 fathoms, 21 species range from 25 to 100 fathoms, and there is essential disappearance of corals at 400 fathoms, but corals have been collected from bottoms as deep as 1,150 fathoms. The deeper water forms are not reef builders (Vaughan).

Corals live within the temperature range of from 30°F. to normal heating of marine waters, but the greatest abundance is in the range from 73 to 78°F., in which range 70 species live about the Hawaiian Islands. Living reef corals can endure a temperature as low as 64°F., but 68°F. is near the lower limit at which they may thrive. Noncolonial forms dwell in much colder waters.

The data given above apply to living forms of Anthozoa and may have application to near relatives of these forms that lived in the Tertiary, but the data certainly should not be considered to have positive application to reef-building corals of the Paleozoic and Mesozoic eras. The evidence of the sediments associated with the ancient reef corals shows that the waters were shallow, but neither the sediments nor the corals permit assumption of tropical temperatures.

The response of different species of corals to sedimentary conditions is illustrated by their distribution on the exposed barrier, the barrier flat, and the lagoon of the Cocos Islands of the East Indies (Vaughan). The exposed barrier has 16 species; 20 species live in pools on the barrier; and 23 species are in the lagoon. The lagoon and the exposed barrier have only 3 species in common, of which one has its exoskeleton so different in the two places that if found in the geologic column they would be considered distinct species. The distance between the lagoon and the exposed barrier is less than ½ mile. Of the 20 species in the pools on the barrier flat, 6 live in the lagoon and 2 on the exposed barrier. The barrier flat does not have a majority of its species living either on the exposed barrier or in the lagoon. If the barrier flat should extend itself either over the lagoon or the exposed barrier, the faunal differences now shown laterally would show in the vertical section. The strata bounding many unconformities in the geologic column have a greater number of common species.

PORIFERA (SPONGES) (DE LAUBENFELS)

Porifera are generally rare in the geologic column, although locally present in abundance. Essentially nothing is known of the factors of environmental control of extinct forms. Living forms dwell in both marine and fresh waters, but marine forms are larger and more abundant than fresh-water forms. They seem to have greatest development in clear and warm waters.

BRYOZOA

Not much seems to be known of the influence of sedimentary conditions on Bryozoa beyond that most of them seem to live in shallow waters of the neritic life zone, prefer clear waters and desire objects on the bottom or floating objects for attachment, which on the bottom may be anything from rocks to plants.

Most of them seem to live in tropical and warm temperate waters, but they are also extremely abundant in extremely cold waters. Many live on shells and they are frequently brought up by fishing nets on the north shores of Newfoundland from waters that are near freezing temperature. Until more is known of the adaptations of living bryozoans to sedimentary conditions, it will not be possible to state the significance of extinct forms with respect to sedimentary conditions.

BRACHIOPODS (COOPER, SCHUCHERT)

Living brachiopods are altogether marine and belong exclusively to the benthos. All facts indicate that they have always been as they are at present. They live in all depths from very shallow to more than 14,000 ft., and it is probable that each species is adapted to certain depth and temperature ranges. Depth ranges of some living species extend through several thousand feet. Most living brachiopods must have bottoms with solid objects to which they may be attached. Attachment is by a fleshy pedicle, and if solid objects are not present and the brachiopod shells show by their structure that the pedicle was there, the probability is that the bottoms were not the places of life. *Lingula* is the only living representative of unattached forms, but there are many extinct forms, as *Productus*, *Chonetes*, and others, that were not fastened to the bottoms by the pedicle, as that structure was not present. *Lingula* lives in muds, and related extinct unattached forms may have lived under like conditions.

Most inarticulate brachiopods seem to live on neritic bottoms, but one living species is known to extend to the depth of 4,400 meters. This wide vertical range of living species does not permit much exactitude in making assumptions respecting depths in which the ancient forms lived. The sediments in most cases prove that the known extinct inarticulate brachiopods lived in shallow waters. Living articulate species also live mostly on neritic bottoms, but some descend to depths of nearly 4,000 meters. This likewise suggests that most of the ancient forms lived on shallow bottoms, but if one relies on the brachiopods themselves for proof, this assumption could not be verified. However, the sediments again prove shallow-water origin.

Shallow-water species of living brachiopods usually have thick shells, whereas those of deep waters are thin, fragile, and often almost transparent. The thick shells are obviously responses to wave violence or to predatory attack. This characteristic probably has the same significance if present in extinct forms. Many ancient brachiopods exhibit broken and healed or repaired shells. This breaking, if due to wave action, indicates shallow waters, but as it cannot be stated that the breaking was not due to predators, nothing is proved respecting depth.

Living brachiopods dwell in both warm and cold waters, with each species adapted to its own temperature range. Species of the same genus may have different temperature ranges. Since this is the case at present, it was probably the case in the past; hence it is impossible to draw any conclusions as to the temperature significance of extinct species.

Certain species of extinct brachiopods are known from the associated sediments to have lived in clear waters, but other closely related species lived on bottoms on which mud was deposited. The associated inorganic sediments afford better evidence respecting turbidity than the brachiopods.

Certain living linguloid brachiopods prefer waters with salinity below normal, and perhaps some extinct brachiopods had the same preference, but again there are no means of ascertaining whether such was the case.

In general, brachiopods indicate marine waters but tell little or nothing with respect to past temperatures. Strong and thick shells suggest homes in the zones of wave violence; thin and fragile shells suggest deep water. More information on the adaptation of living brachiopods to sedimentary conditions is necessary before it will be possible to do much with the extinct brachiopods in reconstruction of ancient sedimentary environments.

ECHINODERMS

Echinoderms are exclusively marine and, so far as is known, have never entered fresh waters. They dwell on all kinds of bottoms of marine waters. Holothurians live mostly on muddy or sandy bottoms and range according to species to depths of 600 meters. Echinoids live on muddy, sandy, and solid bottoms, each species having its particular preference, although there are some species that seem able to live anywhere from wave-beaten shores, where they drill holes in many kinds of rocks, to very quiet waters. Most live in shallow waters, but a few species are known to live to depths of 5,400 meters. Starfish are abundant on modern bottoms and live mostly in shallow waters. The rarity of fossils of extinct starfish renders them nearly valueless for reconstruction of ancient environments. The evidence available indicates that most fossil forms lived in shallow waters. The extinct benthonic crinoids, blastoids, and cystoids also seem to have lived in shallow waters, but living forms of benthonic crinoids are found to depths of 4,200 meters. Crinoids seem to have preferred waters free from much suspended sediment. Benthonic forms probably never lived in very shallow waters where waves were violent, as such would have been too dangerous to them. Living species have considerable temperature range, and it is assumed that the extinct forms probably had species adapted to every temperature in the marine waters of the time. Some living crinoids are free swimming, as were also some of the extinct cystoids and crinoids. Remains of the free-swimming crinoids may become incorporated in the sediments of any kind of bottom, and hence their remains state nothing with respect to the environmental conditions of deposition of the sediments in which they are contained. Free-swimming cystoids of the past are no more significant.

In general, therefore, echinoderms do not offer much aid in reconstruction of the environments of deposition of the sediments in which they occur, beyond the reasonable certainty that the sediments are of marine origin.

GASTROPODS AND PELECYPODS (BARTSCH)

These two phyla are among the most important living aquatic animals. Gastropods live in both fresh and salt waters and also on the land, and some families have both fresh- and salt-water genera. Pelecypods have both fresh- and salt-water representatives. Members of the two phyla have aquatic forms adapted to every depth of the sea and to every temperature. Some pelecypods live on solid bottoms, others on mud and sand bottoms. Some are buried in muds or sands; some dwell in the littoral zone and are exposed at each low tide; and others drill holes in rocks and wood for homes, which ultimately become their tombs. Gastropods are about as versatile. They are found on mud, sand, and solid bottoms of both fresh and salt water, about coral reefs, and in the littoral zone. Some land forms live on plants and others on the ground.

Shells of the bottom-dwelling, shallow-water marine forms of these two phyla tend to be heavy; shells of those that dwell in mud are fragile. Shells of deep-water forms tend to be fragile. Some fresh-water shells are massive, and others are thin and fragile. Shells of land gastropods are thin. Some species of gastropods endure both fresh and salt water, for example, *Littorina*, which ventures into streams, and some species of pelecypods prefer waters below normal salinity. There is no assurance that in the past the two phyla did not have species similarly adapted to a large variety of conditions, and furthermore there is no reason to believe that extinct forms related to modern forms were adapted to the same conditions. Knowledge of the significance of these two phyla must be derived from learning the significance of each extinct species from the materials with which it was associated while living. Certain species have been proved to have lived in marine and others in fresh water, and if the shells of these species are found in the places where they lived, the significance is plain. If shells are thick and massive, a shallow-water origin for the sediments containing the shells may be postulated, provided the shells are where the animals lived, but it is not determined that the waters were fresh or salt. If shells are thin, there is more uncertainty, as thin shells are produced by land gastropods, burrowing and boring pelecypods, stream and lake pelecypods and gastropods, and also deep-water gastropods and pelecypods.

Cephalopods (Miller and Furnish)

All living cephalopods are marine, and they seem always to have lived in the sea. Some extinct shelled cephalopods are supposed to have lived a nektonic life, but it is probable that most were vagrant benthos. Dead shells may have become pseudoplanktonic until they became waterlogged and sank. Nektonic forms are abundant at the present time in the shell-less squids and devilfish. These have internal shells in some of the squids, and the extinct form of *Belemnites* had quite strong internal shells. These shells settle on all marine bottoms and have no meaning with respect to sedimentary conditions. Nektonic and pseudoplanktonic forms with shells likewise left and still leave shells over all bottoms. Benthonic forms have sedimentary significance, but it is not understood, as little is known respecting their life habits and the preferred environments. The quite common association of cephalopods with reef corals suggests preference of cephalopods for warm water, but they are also found in all other kinds of sediments, and it is probable that cephalopods were adapted to every temperature prevailing during their existence. It is also probable that they were adapted to many depths and many kinds of bottom.

It was once postulated that the distribution of ammonites indicated climatic zones, but as knowledge of the distribution has been extended, it has become obvious that the postulate was largely due to ignorance of distribution, and, although there probably were climatic zones, they do not appear to be shown in the distribution of the ammonites.

In this phylum, as in others, the significance of each species must be learned from the associated sediments and organisms.

TRILOBITES (RAYMOND)

Trilobites seem always to have been marine and to have been adapted to a wide range of environment. Some were pelagic plankton, others pelagic nekton, and still others belonged to the benthos. Planktonic adaptation is thought to be shown in a vestigial pygidium and possession of abundant spines. A nektonic adaptation seems to be indicated by thin shells and large pygidia, and the species referred to the benthos have large shells, poor pygidia, and large shovellike cephala.

Planktonic and nektonic forms signify nothing respecting environments of deposition of the sediments in which they are imbedded. Bottom-dwelling forms tell nothing in their structures respecting depth and temperature. This must be determined from the containing sediments. The fact that all trilobites are supposed to have molted and that the molts were light with large surface and hence easily carried by waters of low competency doubtless led to wide distribution and deposition in many environments in which the trilobites did not and perhaps could not have lived. Caution is hence essential in the uses of trilobites for interpretation of sedimentary conditions. Fragments ought never to be used, and reliance should be based on complete specimens.

Other Arthropods (Raymond)

These live in fresh waters, in the sea, and on land. Fossil forms may not even enable one to differentiate between continental and marine sediments, but individual species may be proved to have had one or the other habitat. Most are vagrant benthos, but they molt periodically, and the molts are easily broken and transported to any depth and even carried on shore. Arthropods are favorite food for birds at present and probably were so in the past, and hence the exoskeletons may have been carried far inland as they are at present. Many belong to the plankton and hence may be deposited in waters of any depth. They live at the present time in waters of all temperatures and doubtless did in the past. The mere remains of members of this group of animals have little value in interpretation of past sedimentary conditions.

Barnacles

Barnacles have always lived in the sea, and they are sedentary benthos or epinekton or epiplankton. The pelagic forms may be found anywhere and mean nothing except that the containing strata are probably marine. Most benthonic barnacles dwell in shallow water. They are particularly abundant in the littoral and for a short distance below, but they extend to known depths of 900 meters. The benthonic forms have little meaning unless attached. A marine shallow-water environment should be postulated if they are present in large numbers in any part of the geologic column.

Merostomes

Living merostomes dwell in very shallow marine waters. Extinct forms from the Teritary of Germany and the *Prestwichia* and *Belinurus* from the Carboniferous of North America seem to have lived in fresh water. The ancient eurypteroids have been the subject of much controversy. Some consider them to have lived in streams and to have floated out to sea, and others ascribe a marine habitat to them. Until the habitat has been definitely proved it is vain to place any reliance upon the eurypteroids to prove sedimentary conditions.

VERTEBRATES (CASE)

Most vertebrates are land animals. Aquatic forms belong to the nekton and have their remains spread over every kind of sea bottom from the littoral to the abyssal and over the bottoms of rivers and lakes. In the sea it is probable that the remains are more abundant in deposits of shallow bottoms where burial is rapid, as food seems to be more abundant now in the shallow waters, and probably was the same in the past and the animals would be with the food. However, burial would have to be rapid; otherwise borers and scavengers of the bottom would soon destroy all remains except very resistant structures which on bottoms of slow deposition might be so abundant in sediments as to suggest dense population. The skeletal structures usually give the habitat away, as paddles and fins prove aquatic forms, wings indicate flyers, and little modified legs and feet land animals. The feet indicate whether the animal was a burrowing, a cursorial, a tree dwelling, a saltatory, or a swimming form. The teeth indicate something of the food, and the shape of the foot may give some information respecting the solidity of the surface upon which the animal walked.

A second form of evidence respecting environmental conditions of the extinct vertebrates is based upon the habits and homes of existing animals that are related to the extinct forms. This is based on analogy and may have little value, as shown by the Pleistocene elephants and mastodons that lived amid the ice fields of Alaska and Siberia and the present elephants of the tropics.

Dead land animals when inflated with decomposition gases readily float and thus may be transported downstream and ultimately deposited in stream channels, swamps, lakes, and perhaps the sea. Obviously the places where these are found indicate nothing respecting living conditions. Only when buried in swamps and sand dunes may the structures of the animals be related to the sedimentary conditions under which they lived. In all other cases, except the occasional animal caught in a crevasse in the ice as was the Siberian mammoth, the animals lived in certain places and were buried in others, and thus the fossil organisms state nothing respecting the sedimentary conditions of the habitat except that they were such as to transport the animals away.

RELATIONS OF SEDIMENTS TO ORGANISMS

Most sediments after deposition contain plant and animal matter, and with such food available from the beginning of organic history it naturally follows that organisms would become adapted to this form of food. This could be obtained by large organisms only at the cost of swallowing large quantities of sediments or soils. The earthworms probably take first place in this respect in the soils. The soil materials and the contained organic matter are swallowed, and after extraction of nutriment according to the needs of the animal the unused substances are excreted. This is commonly done on the surface, so that the subsurface materials are progressively conveyed upward. It has been estimated that the annual quantity brought to the surface under favorable conditions amounts to 18 tons per acre.

Animals of various kinds throng many parts of the bottom of the sea, and there is little in the way of organic matter in the underlying sediments accessible to these animals which they have not sought to obtain. The quantity swallowed often exceeds the weight of the animal. *Sipunculus nudus*, a worm living in the Bay of Naples, when filled with sediments was found to weigh 19.08 grams, of which 10.03 grams consisted of sands contained in the alimentary canal (Dapples).

Marine worms eat sediments in the sea just as earthworms do in soil. Among them are the lobworms that live in the muds of shallow water. The work of lobworms has been studied on the coast of Northumberland, England, where it has been estimated that their castings (excremental particles) existing on the surface at any one time average 84,423 per acre or around 50,000,000 per square mile. This is really not large, as it amounts to only about two castings per square foot, a figure that is not particularly impressive but adequate to bring through the intestinal tracts of the worms 2,000 to 3,000 tons per acre annually. As these worms burrow to a depth of about 2 ft. (0.6 meter) this thickness is subjected to a slow but rather thorough mixing, and everything to the depth of burrowing passes through the intestinal tracts in about 22 months. Some of the sediments are eaten several times (Davidson).

The chief gross foods of holothurians are muds and sands swallowed for the contained organic matter, and one writer states that they "continuously shovel sand and anything it may contain into their mouths, whence it passes through the entire alimentary tract of the animal." The excretions consist of solid threads, which are 2 to 5 mm. in diameter. These may be thrown 7 mm. from the animal's body. The first discharge is in the form of an irregular pile which terminates in loosely looped cords in the final stages of ejection (Flettely and Walton).

Echinoids ingest vast quantities of the muds and sands upon which they live, and these make repeated passages through their intestinal tracts. The particles are doubtless abraded in passage, and there must be some solution. Everything on the bottom small enough to be swallowed is taken, and, if the particles are too large for ingestion, many echinoids have strong and sharp teeth to reduce them to a dimension suitable for swallowing. Shell matter is thus destroyed or reduced to such fine particles as to be impossible of recognition (Kindle).

All arthropods are great scavengers. They break shells with their claws, swallow everything offering a morsel of food, and grind all thoroughly in the intestinal tracts, some even having gastric mills for that specific purpose. The materials always emerge in smaller dimensions than when they entered. There can be little doubt that great masses of shells are annually reduced to small fragments in this way.

Mollusks literally stuff themselves with finely divided bottom materials and eject the inorganic materials largely robbed of organic matter.

Many fish feed on the bottom fauna just as animals browse in a pasture. Among such are the cod, hake, and haddock. Entire shells are often swallowed and disgorged unchanged except for some solution. Fish browse over coral reefs of the present and doubtless did so in the past, and it is probable that the ancient crinoid colonies at times suffered horribly from this cause. The great absence of calices in crinoidal deposits is perhaps explainable on the basis that the heads were bitten off by browsing fish.

Another phase of sediment destruction or modification is effected by boring animals. The shipworm and similar bivalves bore holes in everything with which they come in contact, and rock, wood, shells, bones, and any organic deposits may be literally riddled. Bones of aquatic animals are rare over bottoms on which boring mollusks live, in spite of the fact that the animals may be extremely abundant in the overlying waters. The sulphur sponge, *Cliona*, burrows in shells, corals, algae, and limestones when young, and many gastropods drill holes into living mollusks to attack the living prey within. A small annelid, *Leucodora*, drills holes in chalk, sandstone, shale, mica schist, and probably other rocks, and on some seacoasts of France as many as 250,000 to 300,000 worms per square meter have been counted (Dapples). *Gephyrea* riddles the *Lithothamnium* deposits of Funafuti, and on some rock-bround coasts both hard and soft rocks are riddled with holes that are drilled by echinoids, holes in which they ensconce themselves for protection.

After the macroscopic organisms have done their work, the destruction is continued by microscopic animals and plants. Bacteria are probably most important in this respect. Species are thought to be adapted with respect to foods, temperature, depth, light, and aeration to every environment, and unless sediments are rapidly buried there is little or no organic matter that they do not attack. They screen the sediments thoroughly for food. Many products result from their activity and these may extensively modify the sediments.

As long as they contain free oxygen, the organic deposits of swamps are bored through and through by bugs, beetles, worms, and aerobic bacteria. When the oxygen is exhausted or excluded, anaerobic bacteria continue the task. It is only where sediments are rapidly deposited or where organisms produce antiseptic products that organic work is brought to a close and destruction of organic matter is not carried to completion.

Some sedimentary impress is made by large animals. Many animals use pieces of rocks or shells in a part of the alimentary tract for grinding food. Among such are the living herbivorous birds, some living reptiles, and seals (Emery). The rocks in the alimentary tracts of seals and penguins are popularly known as "*ballast.*" The extinct dinosaurs used large quantities of stones for grinding their food, and rocks up to 5 cm. in diameter were swallowed. These are known as *stomach stones* or *gastroliths.* They are occasionally common in deposits containing dinosaur bones. Few gastroliths are composed of soft rock, and they usually are highly polished.

The generalization seems to be valid that few sediments of lakes and seas of slow deposition have escaped passage through the intestinal tracts of organisms and that most slowly deposited sediments are excretions. Buchanan, a prominent student of ocean sediments of nearly 50 years ago, stated as his conviction that "the principal agent in the comminution of the mineral matter found at the bottom of both the deep and shallow seas and oceans is the ground fauna of the sea, which depends for its subsistence on the organic matter which it can extract from the mud" and "the matter forming the bottom of the sea is being continually passed and repassed through the bodies of the numerous tribes of animals which demonstrably subsist on the mud and its contents." The work begins with larger nekton and larger benthos that break or chew the shells, is continued by worms and wormlike animals, which is where Buchanan may have considered it finished, and is finally completed by microorganisms, among which bacteria are preeminent.

Excretions are mostly in the form of pellets 1 mm. or less in diameter. These are aggregates of various organic and inorganic substances held together by some sort of cement provided by the animal. This cement is often so poor that the pellets crumble on ejection. If they are sufficiently coherent and survive later organic action, the pellets flatten under the burden of accumulating sediments, and after lithification they resemble and may be mistaken for oölites. If the particles fall apart, the result is a fine-grained deposit, which, because of burrowing animals, may be totally devoid of stratification (Moore).

Reworking of sediments by organisms may lead to the destruction of all stratification in all slowly deposited fine-grained sediments except those that are deposited beneath waters that are more or less permanently low in oxygen and high in hydrogen sulphide or those that contain little or no organic matter. The organisms that do the reworking cannot thrive under the first exception, and the absence of food in the second does not invite entrance of them. Sediments of rapid deposition cannot be completely reworked.

The final result of organic action upon sediments may so change their composition, appearance, texture, and stratification that they could not be identified as those originally deposited. Thus, sediments on deposition may be composed in large part of organic matter. This may be entirely eliminated, and the product after lithification may have inorganic constituents only (see diagenesis).

PRODUCTION OF SEDIMENTS BY ORGANISMS

Organisms produce various protecting and supporting structures which on the deaths of their builders become sediments. The chief supporting and protective structures are wood, bones, shells, exoskeletons, and tests. Wood is a compound of carbon, oxygen, and hydrogen. Bones are mostly calcium phosphate and calcium carbonate. Some organic structures are composed of calcium and magnesium carbonates. Exoskeletons are composed mostly of chitin or calcium carbonate, the latter often serving as a stiffening for the former. Most tests and shells are either calcium carbonate or silica; a few are composed of other substances. The substances forming the supporting and protecting structures of organisms and their chemical compositions are shown in Table 7.

Table 7. Solid Materials Contained in Organisms and Organic Structures

- Protozoa.
  - Foraminifera: calcite, chitin.
  - Silicoflagellates: opal (?).
  - Radiolaria: opal, acanthin, strontium sulphate, barium sulphate (?).
- Sponges: calcite or aragonite, opal, spongin.
- Coelenterates.
  - Graptolites: seracin.
  - Corals: aragonite and calcite.
- Annelids: calcite, phosphate, chitin.
- Bryozoa: calcite, chitin.
- Brachiopods: calcite, aragonite, chitin, dahllite.
- Mollusks: calcite, aragonite, phosphate, chitin.
- Echinoderms: calcite.
- Arthropods.
  - Crustacea: calcium carbonate and chitin.
  - Insects: chitin.
- Vertebrates: lime carbonate and phosphate, magnesium phosphate, carbonate and chloride of sodium, collophane, keratin.
- Plants.
  - Diatoms: opal.
  - Bacteria: $CaCO_3$, Fe_2O_3, FeS_2, S and other substances are formed.
  - Algae: calcite, aragonite, magnesium carbonate, cellulose.
  - Bryophytes and higher plants: cellulose, resins, waxes, fat, gums.

These various organic structures accumulate in proportions related to the kinds and numbers of the organic population and the destructive agencies that work upon the dead materials. Under some conditions plants pile up vegetable materials in swamps that startle imagination, and shells are produced in several environments with extraordinary rapidity. Much plant matter is produced in seas and lakes. It has been stated that in Danish waters, *Zostera*, an alga, annually produces 4 tons of dry plant matter per square mile (Petersen), and Lake Mendota, a fresh-water lake at Madison, Wisconsin, with an area of about 20 square miles (39 square km.) annually produces from its aquatic plants about 4,850,000 lb. (2,203,000 kg.) of dry plant matter (Rickett). Much of this is as rapidly destroyed by organisms living in the lake, just as in the sea the enormous annual production of plant materials largely disappears.

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CHAPTER VI

TRANSPORTATION AND DEPOSITION OF SEDIMENTS

GENERAL CONSIDERATIONS

A particle becomes a sediment when movement from the place of origin begins. This movement for terrigenous sediments is initiated at the moment of detachment from the parent rock, and it may continue indefinitely for millions of years. It is probable that some material now in solution in the oceans has been there since the first appearance of seas, and some sands of some beaches have moved with them for millions of years. Sediments are subjected to physical and chemical changes during transportation, and many particles acquire features connected with the method of transportation.

Transportation is done by five agents, which may be placed in three classes:

1. Water and atmospheric transportation. Transportation of sediments through traction and suspension in turbulent currents of low viscosity, and, in the case of water, in solution.
2. Glacier and gravity transportation. Transportation of materials through incorporation in masses whose movement resembles lamellar flow in highly viscous fluids.
3. Organic transportation. Transfer of materials connected directly or indirectly with the movement of organisms.

Transportation by water and ice is indirectly due to gravity, but transportation by these two agencies is very different from that in which gravity functions directly. Many sediments have been transported by each of the agents at some time in their history, as may be illustrated by consideration of a single example. Winds are blowing sands and dusts acquired from the floodplain of the La Crosse River near Sparta, Wis., eastward up the valley of Crystal Creek, a small tributary entering the La Crosse River from the east. The sediments are deposited in places over this valley as dunes and dustfalls. These sediments were deposited on the floodplain of the La Crosse River by stream waters that acquired them from the outwash deposits of the Pleistocene glaciers. The glaciers acquired the sediments from the Cambrian sandstones, for whose deposition the sea was responsible. Thus, the sequence of transportation and deposition as far as is known was marine transportation and deposition, ice transportation and melt-water deposition, stream transportation and deposition, and lastly wind transportation and deposition.

Water and the atmosphere are the two chief agents of transportation. Ice is of great importance, but it does not compare quantitatively with the other two agents. The qualitative results of the direct pull of gravity far exceed the importance of the quantity involved, and, moreover, gravity transportation serves largely as a feeder for water transportation. The transportation done by organisms is also a feeder for water and atmospheric transportation.

CLASSIFICATION OF DIMENSIONAL UNITS

Any consideration of transportation and deposition of sediments raises the problem of dimensions of particles and terminology in terms of dimension. Several classifications of dimensional units have been proposed. The one most in use by American sedimentationists is that of Wentworth (1922). This is given below, and his terminology with the dimensional significance as given by him is used in this book.

TABLE 8. CLASSIFICATION OF SEDIMENTARY PARTICLES

Name of Particles	Dimensions, Mm.
Boulder	256 or above
Cobble	64 to 256
Pebble	4 to 64
Granule	2 to 4
Very coarse sand grain	1 to 2
Coarse sand grain	½ to 1
Medium sand grain	¼ to ½
Fine sand grain	⅛ to ¼
Very fine sand grain	1/16 to ⅛
Silt particle	1/256 to 1/16
Clay particle	Smaller than 1/256

Alling has proposed a metric scale of measurement applicable for thin and polished sections of sedimentary rocks where measurement can be taken in only two dimensions. The major divisions of his scale are based on the constant ratio of 10, the minor divisions on the ratio of the fourth root of 10. The scale is given in Table 9.

TABLE 9. METRIC SCALE FOR THIN AND POLISHED SECTIONS

Major divisions	Minor divisions	Size range, mm.	Aggregate term
Boulder...........	Coarse	560 to 1,000	Boulder stone
	Medium	320 to 560	
	Fine	180 to 320	
	Very fine	100 to 180	
Cobble............	Coarse	56 to 100	Cobblestone
	Medium	32 to 56	
	Fine	18 to 32	
	Very fine	10 to 18	
Gravel............	Coarse	5.6 to 10	Gravelstone
	Medium	3.2 to 5.6	
	Fine	1.8 to 3.2	
	Very fine	1.0 to 1.8	
Sand..............	Coarse	0.56 to 1.0	Sandstone
	Medium	0.32 to 0.56	
	Fine	0.18 to 0.32	
	Very fine	0.10 to 0.18	
Silt...............	Coarse	0.056 to 0.10	Silt stone
	Medium	0.032 to 0.056	
	Fine	0.018 to 0.032	
	Very fine	0.010 to 0.018	
Clay..............	Coarse	0.0056 to 0.0010	Claystone
	Medium	0.0032 to 0.0056	
	Fine	0.0018 to 0.0032	
	Very fine	0.0010 to 0.0018	
Colloid...........	Coarse	0.00056 to 0.0010	Colloid stone
	Medium	0.00032 to 0.00056	
	Fine	0.00018 to 0.00032	
	Very fine	0.00010 to 0.00018	

TRANSPORTATION AND DEPOSITION BY WATER

Methods of Transportation

The results of transportation by water depend upon whether the transportation is done in streams or in lakes and the sea. The agents of transportation are the same, but there are differences in the intensities. Transportation in both is accomplished in each of the three ways: *traction*, *suspension*, and *solution*. The agents of deposition are also the same, but vary in intensity. Results in the deposits are very different.

Traction and suspension are physical methods of transportation that depend upon current velocity and turbulent movement for ability to transport all particles above colloidal dimension. Sediments transported by traction are usually termed *bed load*. As velocity changes from one rate to another, some sediments transported by traction may be changed to suspension if the velocity increases, and from suspension to traction if the velocity decreases.

There are two kinds of movement of water of currents: laminar (streamlined) and turbulent. Robertson and Rouse divided these into tranquil and rapid laminar and tranquil and rapid turbulent. In laminar movement the water travels in parallel bands and, according to Hjulström, this form of movement is confined to streams of low velocity and is to be reckoned only in fractions of a millimeter per second and may be found only in shallow watercourses. Robertson and Rouse, however, have shown that laminar movement can also exist under conditions of high velocity. Hjulström states that at a certain critical velocity laminar movement changes to turbulent when there is a considerable range of varied and mixed movement. Turbulence increases with increase in velocity until a maximum of 23.5 meters per second is reached when the water breaks into spray. The velocity to form spray may not be so high. In turbulent movement there are rapid variations in velocity around an average value giving rise to pulsating movements of the water and also creating eddies and rollers (Leighly, O'Brien).

Turbulence decreases from a maximum at some depth in a current to zero at the surface and the bed of the stream. Any thread of maximum velocity has a bordering area of maximum turbulence on each side of the vertical line in which the maximum velocity lies, and each thread of maximum velocity in a broad stream is accompanied by two threads of maximum turbulence, as shown in Fig. 23 from Leighly (1934). A wide channel having several threads of maximum velocity does not possess the relative stability of a channel whose depths and width are such that there is only a single thread of maximum velocity. *Austausch* or *Austausch coefficient* is a term used as a measure of turbulent mixing, and it is defined as the product of mass and transverse distance traveled in unit time by the fluid in turbulent motion passing through a unit area that is conceived as lying parallel to the general direction of flow. It corresponds to the English term of *eddy conductivity.*

Transportation by water largely depends upon turbulence in the water among the fragments forming the bed over which currents move of an intensity at least sufficient to loosen the fragments, so that the forward thrust of the water may set them in motion. Places of most intense turbulence move more material upward from the bed than elsewhere, thus making the water at those places turbid. There is, of course, interaction between the water and the bed, and this leads to constant adjustment of the suspended load to the irregularities of the bottom. Local intensification of turbulence above some minimum determined by the nature of the material forming the bottom increases the quantity of sediment moved from the bottom, whereas diminished turbulence leads to deposition. Under conditions of constant discharge an equilibrium is ultimately attained; after which the bottom is not further affected as long as conditions remain unchanged.

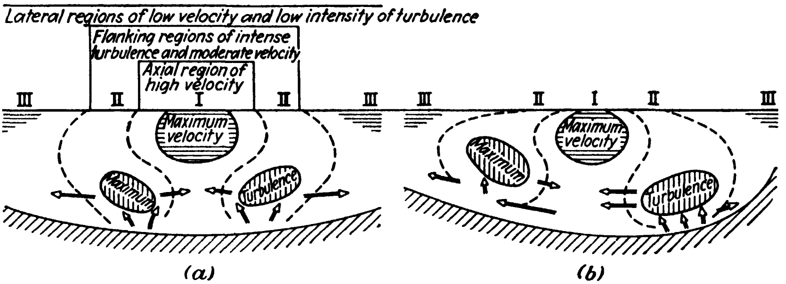
Lateral regions of low velocity and low intensity of turbulence
Flanking regions of intense turbulence and moderate velocity
Axial region of high velocity
III
II
I
II
III
III
II
I
II
III
Maximum velocity
Maximum
Turbulence
Maximum velocity
Maximum
Turbulence
(a)
(b)

FIG. 23. Diagram showing relation between maxima of turbulence and of velocity in cross sections of symmetrical (*a*) and asymmetrical (*b*) transverse profiles. Principal directions of diffusion of suspended matter are indicated by arrows. The broken lines show approximate boundaries of regions I, II, and III. (*After J. B. Leighly, Geol. Rev.*, 24 (1934), 463.)

Horizontal gradients of turbidity are set up in currents, and the suspended matter diffuses horizontally in the directions of lesser turbidity. It thus enters into regions of less turbulence. If velocities here are higher than in the places of derivation of the suspended sediments, they are carried downstream; but, if the velocities are less, the sediments must settle. Regions III of Fig. 23 after Leighly are those into which turbidity migrates. These are of indefinite extent laterally and include all areas of quiet water continuous with regions II. These are the regions of deposition of suspended matter diffused from regions II, the regions of maximum turbulence and turbidity. Region I of maximum velocity carries down-current any suspended sediments that diffuse therein. Leighly (1934) summarizes the teachings of Fig. 23 in the statement that "All stream deposits are formed in areas that are lateral to some thread of maximum velocity and its flanking areas of maximum turbulence."

Water increases in weight to the extent that fine sediments are in suspension, and if water so provided with sediments is in contact with water without or with a smaller suspended load, the heavier water moves as a density current in the direction of, and below, the lighter water. In an ordinary stream it is the water that moves the suspended sediments, but in a density current it is the weight of the suspended sediments that is the driving force propelling the water. Density currents may be seen where a stream brings muddy water to a lake or reservoir, and this muddy water as a density current sinks beneath the clearer water to flow for some distance, or on the shore of a lake where wash of waves places material in suspension. The containing water flows as a density current toward the bottom of the lake. Immense quantities of fine sediments are thus transported over bottoms of standing water. It has been found in Lake Mead in the Colorado Canyon that 77 per cent of the particles in density currents were finer than 5 μ and 97.7 per cent were smaller than 20 μ. Particles larger than 50 μ were not present (Bell).

Particles transported by traction do so by rolling, sliding, or making small jumps. Jumping is termed *saltation,* and for a short period the particles are in suspension. Saltation is favored by irregularities of the bottom, as these favor turbulence, and the irregularities from time to time place rolling or sliding particles in positions from which jumping is easy. Sliding is usually done by particles having two large dimensions and one small. These may slide along with occasional turning over if the dimensions are not too great. At some positions in the turning, currents are directed against their lower surfaces and this may lead to saltation. Positions of stability of particles of these shapes are those in which the down-current side is lower than the up-current side, as in these positions the currents strike the upper surface and thus have no lifting ability. Particles attaining these positions are said to imbricate up-current (Fig. 24). If particles of disklike shapes are placed in any position on a sand bottom and the currents are too weak to move the particles but strong enough to move the sands, these are dug out beneath the particles until they acquire positions inclined up-current. If shapes and dimensions of particles approximate those of a small coin and the specific gravities are not too high, the particles may roll on edge, particularly if velocities are fairly strong and bottoms are smooth and firm (Owens). Spheroidal and ellipsoidal particles naturally roll much more easily than those of other shapes, and the tendency in stream transportation seems to be to reduce particles to these shapes. Particles with shapes of prolate spheroids rotate parallel to long axes, and ultimately such particles rolling on sands come to rest with long axes perpendicular to the direction of current flow. Particles of irregular shapes, if placed on sand bottoms in currents too weak to move them, would have the sands removed from beneath them until they rotate so as to place the longest axis perpendicular to current flow and the particles in such position that they are inclined up-current.

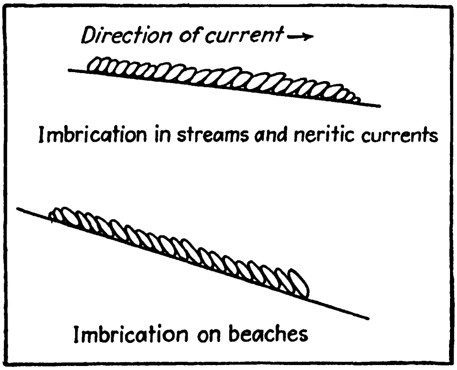

FIG. 24. Diagram showing imbrication on beaches and in streams. The direction of the current in stream imbrication is indicated by the arrow. The current on the beach moves both up and down, the former generally the stronger. If the latter should be the stronger, the inclinations of the rock particles would be into the beach.

Particles moved by traction may travel individually or collectively. Collective traveling is done by sands, gravels, and occasionally larger particles, and it is really a matter of mass travel through individuals of the mass traveling a little at a time. Mass or collective travel is in the form of small, dunelike bodies that have gentle up-current slopes and much steeper down-current slopes. These dunelike bodies migrate down-current by sands moving from the up-current side, which is being eroded, to the down-current side, which is thus extended down-current. According to Hjulström, unaltered forms of "dunes" signify equilibrium without erosion or deposition and perhaps only transportation. This has been termed the *dune phase of traction* (Gilbert). The dune phase develops when the tractional load and velocity are small. An increase in load or velocity brings about disappearance of the dunes, and the sands move as a sheet with gradually increasing density from the surface downward, so that there is no sharp plane of demarcation between the sands in motion and those that are stationary. This is termed the *smooth phase of traction* (Gilbert). Hjulström states that this phase develops when the bottom is being eroded. If deposition takes place when transportation is at the smooth phase, the stratification is not cross-laminated. The smooth phase is transitional to the *antidune phase* (Gilbert), which develops in highly loaded streams from the smooth phase on increase in velocity. In the antidune phase, ridgelike structures develop that are eroded on the down-current side and receive deposits on the up-current side. These have also been termed sand waves. The sand waves or antidunes move up-current as the individual sands move down-current. Antidunes seem to have more symmetrical slopes than dunes and they may also be much higher. In the San Juan River of Arizona they have been observed to be 5 to 6 meters from crest to crest and 0.9 meter from crest to trough. Some have been seen with a height of 1.8 meters. Enormous quantities of sediments are moved in this form of traction (Pierce).

Whether one or other phase of traction obtains seems to depend upon velocity, load, depth, and resistance of beds to current. According to Gilbert, dunes develop if the bed load is small, and these gradually travel downstream. If the velocity increases, the dunes disappear and the smooth phase develops. Either phase may pass into the antidune phase. Both the smooth and the antidune phases seem to have but short duration. The antidunes seem to have their development favored by irregularities on the bottom, and high velocities seem to be necessary. As observed in small brooks on the north side of the Gulf of St. Lawrence near St. John River, the formation of the antidunes brings about a damming of the current, and the water gradually develops a head that ultimately washes off the tops of the antidunes, and the sands are then transported in the smooth phase. This builds broad deposits; again the water is checked; cutting begins at the lower end of a deposit; sinuosity of currents follows; and the transportation passes to the antidune phase. Pierce's description as seen in the San Juan River is as follows: "At one moment the stream is running smoothly for several hundred yards. Then suddenly six to ten waves appear, reach full size in a few seconds, and then as suddenly disappear."

Suspension transportation floats particles, and, except for those of colloidal dimension, transportation depends upon velocity and turbulence. Suspension transportation is aided by irregularities of the bottom, as these produce turbulence, which maintains particles in suspension. Owing to the fact that particles readily change from traction to suspension transportation on increase of velocity and from suspension to traction transportation on decrease of velocity, it is not easy to consider these two methods of transportation separately so far as noncolloidal matter is concerned.

All kinds of material are carried in solution in water, and this form of transportation is independent of velocity.

Transportation of sediments involves consideration of *competency*, *capacity*, and *load* (Gilbert. Other studies have been made by Schaffernak; Mavis, Ho, and Tu; and Mavis, Liu, and Soucek).

Competency is defined as the ability of currents to transport in terms of dimensions of particles. Competency depends upon velocity and turbulence. Both, in turn, depend upon the shape of the channel, irregularities of the bottom, slope, depth, and discharge. A stream has a competent slope for each variation of channel and dimensions of particles of every specific gravity. Other factors remaining equal, there is a competent velocity and also a competent discharge for each dimension of sediments of a given specific gravity. The point intended to be made is that each of the factors upon which competency depends is a variable and for each combination of these variables there is a competent dimension. Competency varies about the fifth power of the velocity.

Currents require higher competencies to lift sediments above the bottom than to transport them after they are lifted. *Traction transportation for a given velocity has greater competency over a smooth surface, and particles in depressions are less readily moved than if the particles are on a smooth bottom.* Particles that form an even surface are moved with difficulty by velocities readily able to transport them if they could be detached from the support of each other. Frequently there are larger particles in the midst of smaller ones that project above the surface. These may be moved, whereas the smaller particles remain stationary. This may be readily observed on bottoms of natural and artificial currents on which large particles roll over a surface of loose stationary sands of much smaller dimension. If at a certain velocity sands or particles of other dimension are transported and the velocity is increased so that larger particles can be transported, these would roll faster than the others if once started.

General conclusions relating to competency as applied to water are as follows: (*a*) for a given kind of sediment, competent velocity must increase as the mass increases; (*b*) with shapes and sizes remaining constant, competency must increase with specific gravity; (*c*) for objects in motion the rate of travel increases with the velocity of the current; and (*d*) competency required to transport particles by traction increases as shapes depart from a sphere or ellipsoid, and competency required to transport particles in suspension increases as they approach these shapes. Specific gravity is extremely important in transportation in suspension—except for very small particles—but the importance is not so great in traction transportation.

A number of studies of competency have been made, among which that of Suchier is perhaps the best. Suchier's observations are given in Tables 10 and 11, which show that a greater competency is required to start motion of particles than to continue it.

The rapid increase in competency with increase in velocity explains the transportation in floods and mountain torrents of blocks weighing many tons.

Capacity is defined as the ability of a current to transport in terms of quantity. Capacity in water depends upon velocity,

TABLE 10. COMPETENCY IN TERMS OF VELOCITY

	Meters per Second
A. Stream bed covered with fine sediment:	
Under action of current alone, no movement with bottom velocity at	0.694
After stirring, movement of sediment for fragments of size of bran began when bottom velocity reached	0.897
Moved particles of size of hazelnuts when bottom velocity reached	0.923
Moved particles of size of walnuts when bottom velocity reached	1.062
Moved particles of size of pigeon eggs when bottom velocity reached	1.123
B. No sediment on stream bed:	
Smallest particles when bottom velocity reached.	1.180
Particles of pea and hazelnut size when bottom velocity reached	1.247
Particles of walnut size are moved when bottom velocity is at	1.476
Particles with weight of 1,000 grams rolled at velocity of	1.589
C. General movement of pebbles:	
To size of pigeon egg at	1.800
To size of chicken egg at	1.717
Particles of less than 2,500 grams weight at	1.800
All particles moved at	2.063

TABLE 11. VELOCITY FOR BEGINNING TRANSPORTATION

Dimensions of particles	Velocity required to start motion m. per sec	Velocity required after starting m. per sec.
Hazelnut size	1.35	0.923
Walnut size	1.39	1.062
Pigeon-egg size	1.45	1.123

turbulence, slope, discharge, depth, shape and dimensions of channel, specific gravity and dimensions of sediments, ratio of depth to width of stream, and degree of mixing of sediments. Some of these factors are not independent, as, for instance, velocity is conditioned by slope, discharge, and depth; depth is modified by discharge and velocity, etc. The ratio of capacity to increase in velocity ranges between the 3.2 and the 4th power of the increase. With constant gradient but variable discharge, the capacity varies as the 3.2 power of the velocity; with constant discharge but slope and velocity variable, the capacity varies approximately with the 4th power of the velocity; and with constant depth but slope and discharge variable, the capacity varies on the average about the 3.7 power of the velocity (Gilbert). Capacity is less for materials of uniform dimension than for materials of several dimensions; this is particularly true of traction capacity. An addition of fine-grained sediments to a certain extent increases capacity for coarser sediments, but a limit is attained when fine sediments bind coarse and prevent initiation of transport.

Suspension capacity is increased if particles have irregular shapes, and is decreased if the shapes are spheroidal. Traction capacity is increased to the extent the shapes of the particles approach spheres or ellipsoids in shape.

The *load* of a current is the actual quantity transported. The load may equal the capacity but commonly does not do so. A load increases the mass of a current. This for a given velocity adds to its momentum. Transportation of a load requires energy, and at the same time the mass is rendered more immobile. It seems probable that currents on the average move more slowly after acquirement of a load than before, unless the water with load is in contact with water without load or a smaller load.

Causes of Deposition

Sediments transported by traction are deposited in direct ratio to decrease in competency and capacity occasioned by decrease in velocity of currents, and traction transportation becomes zero and the sediments thus transported are entirely deposited when velocity falls to that figure, the larger, less spherical, and heavier particles coming to rest in direct ratio to decrease in velocity. The larger particles in suspension behave similarly, so that particles transported by traction in adjustment to a certain capacity and competency may be deposited together with particles transported in suspension in accordance with the competency and capacity occasioned by the same velocity. This places together the largest and most angular particles transported by traction for a given velocity and the largest and most spheroidal particles transported by suspension for the same velocity, although there is a probability that considerable parts of the sediments dropped from suspension become a part of the tractional burden. The mixture of sediments deposited from traction and suspension depends upon the extent to which the two methods of transportation are loaded to capacity and the extent to which the dropped suspended load becomes a part of the tractional load.

Velocity becomes increasingly less important for particles of silt and smaller dimension, as the resistance of the water to particle settling increases with decrease in dimension, and slight turbulence suffices to maintain sediments of small dimension in suspension, and some turbulence is generally present in natural surface waters. Particles of colloidal dimension are independent of velocity and are so small that they do not settle in response to gravity and hence remain indefinitely in suspension, or until several unite to make a larger particle. This is termed *flocculation* or *coagulation*, and the union is brought about by substances in solution, termed electrolytes, and also by colloids carrying electric charges of opposite sign from those coagulated (Boswell). Small additions of colloids of opposite sign (Searle) to a colloidal suspension produce no precipitation, but flocculation is initiated as the quantity added is increased, and further additions bring about rapid precipitation, but an excess of colloid of opposite sign prevents precipitation. Electrolytes also carry electric charges. If these are of opposite sign from colloids in suspension, flocculation and precipitation result. Flocculating ability is related to the presence of ions of opposite electric charge, valency, modifying action of ions carrying same charge as colloids, adsorptions of ions, and the concentration of electrolytes in the solution.

The charge carried by a colloid depends upon the colloid itself and the medium in which it is suspended. Many oxides and hydroxides and colloids of readily oxidizable minerals carry positive charges, whereas minerals not readily oxidized have negative charges. Silica, clay minerals, and humus carry negative charges, but alumina, ferric oxide and hydroxide, lime, magnesia, and hydroxides of chromium, zirconium, titanium, and others are positively charged. Basic dyes carry positive charges, acid dyes negative charges. Flocculation is hastened by the presence of acids and calcium oxide and hydroxide.

The quantity of electrolytes required for flocculation is larger to the extent that the colloids are smaller. Silts and most insoluble substances are flocculated most readily by calcium salts in neutral solution; clay minerals are more readily flocculated in acid solutions.

Searle gives the flocculent ability of some salts as shown in Table 12.

TABLE 12

Hydrochloric acid	30
Nitric acid	28
Sulphuric acid	20
Calcium chloride	15
Calcium nitrate	10
Calcium sulphate	5
Potassium chloride	3
Potassium nitrate	2
Potassium sulphate	1
Sodium chloride	1
Sodium nitrate	1
Sodium sulphate	0.5

The same salt may behave differently with different substances. Thus, calcium nitrate is a better flocculent of silt than calcium hydroxide, but a poorer one for clay in an alkaline solution. Flocculated substances may be *deflocculated*, or *peptized*, by addition of suitable salts followed by agitation. Some of the best substances for this purpose are sodium carbonate, sodium oxalate, sodium citrate, and, if calcium and magnesium carbonates are not present, sodium and ammonium hydroxides.

Colloids may be protected against flocculation through addition of certain colloids or salts. The colloids are termed protective or stabilizing. Thus, silica is stabilized if small traces of potassium, sodium, or chlorine are present, and ferric hydroxide by a trace of ferric chloride. Removal of these substances renders the colloid unstable and subject to flocculation. If a stable colloid and an unstable one in contact are acted upon by an electrolyte that would coagulate the unstable colloid, the stable colloid would function as a protector or stabilizer and prevent precipitation of the unstable colloid, and thus the two may be transported without further flocculation. A common protective colloid in natural waters is the black colloidal organic matter of peat, swamp waters, and soils.

RATES OF SETTLING OF SEDIMENTS

Sediments settle in water and air in accordance with dimensions, shapes, specific gravity of particles, and specific gravity of the water, which in turn is determined by the materials in solution and the temperature. In the case of particles of dimensions at least greater than fine sand, the particles settle almost entirely under the influence of gravity, and other factors are of little importance. Colloidal particles do not settle at all. Between these two limits of particle dimension are those upon which the properties of the medium exert a large influence and determine the rate of settling. To these particles the Stokes equation applies. This is as follows:

$$V = \frac{2}{9} gr^2 \frac{(s - s')}{c}$$

in which V = the rate of settling.
g = the acceleration due to gravity.
r = the radius of the particle.
s = the specific gravity of the particle.
s' = the specific gravity of the liquid.
c = the viscosity coefficient of the liquid.

The formula assumes that the particles are spheres, which is rarely the case. Odén introduced the concept of the effective radius or the radius of a sphere that would settle in the fluid at the same rate as the given particle. c is not constant, but varies with the temperature, as shown in Table 13 below for water.

TABLE 13. RELATIONS OF VISCOSITY TO TEMPERATURE

Temperature, °C.	Centipoise
0	1.7921
4	1.5674
10	1.3097
20	1.0050
30	0.8007
40	0.6560
50	0.5494
60	0.4688

The formula is not applicable to particles with diameter greater than 0.05 mm. Oseen has developed a formula said to be applicable to all particles. According to Lane (1938) the rate of settling of particles larger than 0.3 mm. varies as the square root of the diameter, and below the dimension of 0.1 mm. it varies as the square of the diameter.

TABLE 14. RELATIONS OF DIAMETERS OF PARTICLES TO RATE OF SETTLING

Diameters of Particles, Mm.	Rate of Settling, Mm. per Sec. Temp. 15°C.
1.00	100.0
0.80	83.0
0.60	63.0
0.40	42.0
0.20	21.0
0.10	8.0
0.08	6.0
0.06	3.8
0.04	2.1
0.02	0.62
0.01	0.154
0.008	0.098
0.006	0.055
0.004	0.0247
0.002	0.0062
0.001	0.00154
0.0001	0.000154

The values of temperature above 40°C. have little or no application to natural conditions, but even within the range of natural conditions it is obvious that values of c are important. Temperature also affects the density of the water so that sediments settle more rapidly in warm than in cold water. Stokes's formula also shows the extent to which materials in solution inversely affect the rate of settling.

Experimental work to determine the actual rate of settling has been done by several students. Hazen's results from actual determination, from a curve based on the results of observation, and from the Stokes formula are as shown in Table 14.

SORTING

The three methods of physical transportation—traction, suspension, and colloidal suspension—separate the sediments into three grades and also may separate these three grades from the sediments carried in solution. Colloidal deposition depends upon the action of electrolytes and colloids of opposite sign and separates the fine materials from the sediments carried in suspension and by traction. The colloids may settle with, or separately from, the sediments carried by traction and in suspension. Sorting of the coarse materials is conditioned by decreases in velocity of the carrying currents and is rarely perfect. If velocity decreases rapidly, the loads of both methods of transportation are dropped in a short time and over a limited area. Poor sorting results. If the rate of a decrease in velocity is small, sediments are distributed over a longer space and sorting is better. Loads are dropped from both methods of transportation on the basis of size, shape, and specific gravity; the largest, roundest, and heaviest in suspension transportation and the largest, heaviest, and least round in traction transportation tending to be placed together, with specific gravity not particularly important in traction and very important in suspension. Materials of two methods of transportation are thus deposited together. Excellent sorting in water deposits indicates long and often repeated transportation by waters of different competencies. Loads are moderate when the sorting is good. Large loads produce too large deposition on any decrease in velocity and the capturing of many small particles by larger ones. However, there is one factor in moderate loads that must not be forgotten, that is, under some conditions large isolated particles are more readily transported than small, and these are carried currentward and deposited in the midst of smaller particles. Sorting done in rivers is generally poor. The best sorting done by water is on beaches that pass into the sea over a slowly but gradually deepening bottom. Here the sands are daily and hourly washed by the waves, and, if the shore line is such as to ensure maximum handling and not too great variations in competencies of the waves and currents at the place, the repeated washing of the waves tends to leave on the beach everything the waves bring in that they cannot take out. If the beach is composed of sediments of different specific gravities, the waves would arrange these in bands with the heaviest mineral at the top of the beach and then with decreasing specific gravities to the light at the water's edge (Rubey, 1933). Common beach sands are quartz, garnet, and magnetite (ilmenite). The quartz is low on the beach slope, and then in order upward are garnet and magnetite, and if plant matter is present it may hold the highest place on the beach owing to difficulty in rolling. It seems probable that the highest degree of sorting of fine sediments takes place in lakes, particularly those receiving sediments from glaciers. It does not seem that anything could be more perfectly sorted than the fine sediments that form the winter part of a glacial lake varve.

Physical Effects of Aqueous Transportation

Transportation by suspension does little to the particles transported, as they are too infrequently brought into contact, but there may be a little abrasion and, if currents are sufficiently strong, some impact. The combined effects of traction and suspension result in rounding of particles transported. If particles are very small, less than $\frac{1}{10}$ mm. in diameter, that is, the fine sand or smaller dimension, they usually travel in suspension and never come in contact without a cushion of water between, so that abrasion is negligible. Aqueous rounding in these small particles is entirely the result of solution or perhaps rounding in the intestinal tracts of organisms. Particles larger than sand are rounded through the combined work of abrasion and impact. These strike hard blows on each other, and thus the larger ones have chattermarks that are proportional to the size, texture, hardness, and brittleness of the composing materials. All large particles have ground-glass or mat surfaces. The development of rounding is directly dependent upon size, specific gravity, and distance traveled, and inversely dependent upon hardness. This is expressed in the Mackie formula, which is as follows:

$$\text{Roundness} \propto \frac{\text{size} \times \text{specific gravity} \times \text{distance traveled}}{\text{hardness}}$$

If it is assumed that the initial shape of the particles is a cube, the formula becomes:

$$R \propto \frac{L^3 \times S.G. \times \frac{D}{4L}}{H}$$

or

$$R \propto \text{as} \frac{L^2 \times S.G. \times D}{4H}$$

where L equals the length of one edge of the cube.

It is the author's considered opinion that rounding of sand grains below the dimension of 1 mm. involves extremely long transportation no matter whether the sands are transported by streams or along and on beaches. Observations of sands brought to the sea by the Columbia and other rivers in Oregon showed essentially no rounding after several hundred miles of travel and quartz sands on the beaches and in the coastal dunes of Oregon and Washington showed very little and oftentimes no rounding, and some of these sands must have moved along the beach for many centuries. Moreover, many of the sands are of the second generation in that they were derived from sandstone formations in the coastal cliffs. To produce well-rounded sand grains composed of quartz and feldspars of dimensions less than 1 mm. in diameter requires not a few hundred but many thousands of miles of travel. Russell and Taylor found that the sands transported by the Mississippi River to its lower reaches were more angular than those of the upper reaches. This was interpreted as due to breaking in transportation. This may be the case, but it may also be due to selective transportation.

If particles are moved over bedrock surfaces or over rock fragments, the surfaces become polished and facets are cut on up-current sides of the fragments.

Rounded rock fragments have shapes determined very largely by the structure and manner of breaking of the terrane from which they were derived. Most rock fragments have two long dimensions and one short. These almost invariably form disk-shaped particles. If rock fragments have one long dimension and two short, ellipsoidal particles develop; and if the three dimensions are approximately the same, a spheroidal particle forms. The often-repeated statement that beach particles are disk-shaped is no more true there than elsewhere, if one is dealing with long-traveled particles, as spheroidal and ellipsoidal particles are as abundant on some beaches as other kinds, and disk-shaped particles are by no means uncommon in river and outwash gravels. The local abundance of disk-shaped particles on some beaches is due either to selective removal of the spheroidal and ellipsoidal individuals, because of the greater competency of the currents for particles of these shapes, or to a source in the vicinity that supplies rock fragments of one short and two long dimensions (Wentworth, 1922).

Transportation and Deposition by Streams

The initiation of stream transportation is found in the multiplicity of small temporary rills that develop with the falling of rain or the melting of snow. Every irregularity of slope, every growth of plants, and all plant litter on the surface serve to dam the small rills and result in deposition of parts of the loads they carry. Soil absorption takes toll of the water, and this also forces deposition. The result is that each rill is short and each ends by spreading over deposits of its own making, which in appearance resemble miniature alluvial fans. New rills develop below each such deposit. The materials of the small fans are more or less poorly stratified and sorted. The rill waters may in places be only slightly turbid, and at other places the liquid may be a thick souplike mixture, which is really a highly liquid mudflow. The upstream end of a rill may end in a miniature cliff. These are best developed in cultivated fields and in arid and semiarid regions, but may be found on any slope not well protected by vegetation. Large pieces of rocks on slopes where these cliffs form are undermined and roll down the slopes for short distances. This undermining is often repeated, and thus many large masses of rocks are rolled to bottoms of slopes. The materials of the miniature alluvial fans are ultimately moved, and much finds temporary lodgement at the foot of steep slopes.

TRACTION AND SUSPENSION IN STREAMS

With minor exceptions transportation by traction and suspension in streams is downstream. The frequent changes in velocity and likewise in turbulence lead to equally frequent changes in transportation from traction to suspension and from suspension to traction.

All streams are crooked, largely because of varying resistance to erosion of banks and bed. Streams flowing through alluvial deposits are usually more crooked than those flowing in rockbound channels. The latter streams are actively eroding and are of little interest to sedimentationists, as they form no permanent deposits. They are not further considered. The former are making more or less permanent deposits and hence are of interest.

Crooked streams alternately impinge on opposite banks, and the threads of maximum velocities follow serpentine courses that are more crooked than the channels in which they flow. This fact tends to make the channels and thus the currents more crooked. The thread of greatest velocity is on the convex side of the current and is thus directed against the concave bank. It crosses the channel between concave banks. The impingement of the current against the concave bank raises the surface of the water against that bank, which Gilbert explains is cared for by a downward and lateral movement toward the opposite side of the channel. Some of the energy is directed downward, and if this reaches the bottom, erosion may take place there. The water is stated to acquire a somewhat rotary motion with the axis of rotation parallel to the direction of the channel at the place with the water on the top of the current directed against the concave bank, which is eroded, then a downward movement on that side and on the bottom a movement obliquely toward the concave side of the current. These conflicts of currents lead to turbulence, thus increasing the competency and capacity of the water and erosion of the bottom if the turbulence attains thereto. The current moving underneath toward the concave side encounters resistance and leads to deposition on that side. Thus, the convex side of the bottom is extended into the channel as the concave side is eroded. Leighly (1934) explains the processes differently (Fig. 23). The threads of maximum velocity are on the convex side of the current and are thus directed against the concave bank. On each side of the thread of maximum velocity is an area of maximum turbulence and turbidity. That on the convex side of the thread of maximum velocity encounters opposition to diffusion parallel to the profile of the concave or steeper bank, but, on the concave side of the current, material diffuses without difficulty from the column of maximum turbidity to the waters of lesser turbidity against the convex bank after migrating obliquely across the bottom, and as the waters are quieter and less turbulent there, settling takes place.

As a current crosses from one concave bank to the one next below there are some factors that tend to modify turbulence with the result that there may be either losses or gains in capacity and competency. This may produce deposition and building up, or erosion and lowering, of the bottom. The erosion against and at the base of the concave bank forms deeps; deposition over the crossing produces shoals.

At low water and low velocity, the currents impinge very gently against the concave banks, there is little turbulence and little agitation of the water in the bottoms of the deeps, so that these contain waters that are essentially quiescent. As the currents pass over a shoal immediately above a deep, they may acquire some suspended and some tractional load, both probably of fine materials. The tractional load is rolled into the next deep, and some of the suspended load may settle into the quiescent waters in passage of the currents over it. The currents pass onward to the next shoal with no tractional load and a reduced suspended load. They are thus in a position to acquire these from the shoal. The loads thus acquired are partly lost in the next deep, and so on to the end of the stream. The generalization is thus made that in low water the deeps acquire fine sediments and the shoals lose fine sediments.

At high water, bed velocities become high and diverse, there is much turbulence, and all materials within the range of capacity and competency are moved out of the deeps, materials beyond the range of competency remain. As the currents pass over the shoals energy is dissipated because of much contact with the bottom, and there is less turbulence and deposition may take place on the shoals. Flood waters subside in time, the waters in the deeps become quiescent, and the places of deposition and erosion interchange. The contact between the deposits made in the deeps at low water and the sediments below is an erosional one that is marked by coarser sediments below than above. The coarse material is a part of the underlying and not the overlying unit. If the coarse materials are of gravel dimension, they form not a *basal*, but a *top conglomerate.* The contact between the sediments made on the shoals at high water is also separated from the unit below by a surface of erosion with the coarser sediments in the two units at the top of the lower unit. If the coarse materials are gravels, they form a *top conglomerate.*

As deeps migrate laterally in the direction of the concave bank through erosion of that bank and progressive filling takes place on the concave bank, it results that some remnants of the many unconformities developed by the alternate flood and low waters are preserved.

A profile down a stream channel following the path of the channel would show an alternation of deposits in deeps separated by deposits on shoals. The vertical distance separating a deposit on a shoal from its approximate equivalent in a deep would range from a few to many meters. An example from the Missouri River in Nebraska showed a maximum vertical distance between the two levels of 15 meters, and Todd has shown that in this river a deep had maximum variations in depth, between July 28 and Nov. 18, 1882, of 9 meters. A profile would show alternations of coarse to fine clastics separated by numerous surfaces due to erosion. The deposits of the shoals and deeps would be the same age, and the unconformities, no matter how striking, have little significance.

The materials transported by traction are the coarsest carried by a stream, and they range from silts carried at times of low velocity to gravels and larger particles carried at flood time. Sands may be transported in the form of dune, antidune, and smooth phases of traction, and the deposits may be evenly bedded or be cross-laminated and have current ripple mark.

The sediments transported in suspension ordinarily range from clays to sands, but under conditions of high velocity even granules and pebbles may be transported in this way. There is often much plant and in some instances much animal matter. Stream waters often contain foam, and this readily transports sand, granules, and in some cases pebbles. Sands and pebbles may also be floated on top of the water, particularly if they are covered with an oily or other water-repellent film so as not to be wetted (Jamison). The quantity thus transported may be large (McKelvey, Emery, Evans, Hennessy).

Representative analyses of the suspended load of the Colorado River at the intake of the Imperial Canal at low and high water, 1925, are given in Table 15 (Sykes, 131).

Long transportation by traction and suspension continues chemical decay and leads to physical wear by impact, abrasion, and grinding, particularly on particles transported by traction. The results are that large particles are made smaller and rounder and much material is reduced to minute dimensions. This naturally decreases the tractional load, where most of the wear takes place, and may add to the suspended load. It thus results that sediments reaching deltas and the sea, via long rivers, contain little material larger than sands. Short transportation usually derives materials from limited areas, and the composition is thus likely to be simple, whereas long transportation gives opportunity for materials from many areas to be mingled with probable complexity of lithic and mineral assemblages. However, it should not be forgotten that, if transportation is very long, decomposition may reduce most sediments to clay minerals, iron hydroxides, quartz, and carbonates, the last disappearing in solution.

TABLE 15. REPRESENTATIVE SUSPENDED LOADS OF THE COLORADO RIVER AT INTAKE OF IMPERIAL CANAL

Diameter of particles, mm.	Low water, Feb. 28, per cent	High water, June 8, per cent
2 to 1	1.34	0.22
1 to 0.5	0.89	0.32
0.5 to 0.25	1.94	0.39
0.25 to 0.10	3.73	5.64
0.10 to 0.05	48.00	56.90
0.05 to 0.04	1.41	0.00
0.04 to 0.03	3.20	0.00
0.03 to 0.02	2.07	7.28
0.02 to 0.01	6.14	13.75
0.01 to 0.005	22.42	7.91
0.005 to 0.004	3.52	2.91
0.004 to 0.003	2.21	2.06
0.003 to 0.000	3.15	2.60

Noteworthy is the fact that approximately one-half the load lies between the dimensions of 0.10 and 0.05 mm.

LOADS OF STREAMS

It is difficult even to approximate the tractional load of a stream, and this has been attempted only a few times. Humphreys and Abbot state that the tractional load across a bar in the Mississippi River was 11 per cent of the suspended load, and it was estimated by Gilbert that the tractional load in the Yuba River of California was approximately equal to the suspended load. Collet determined the ratio of bed to suspended load in the Rhone River near Villeneuve as 1:7, in the Linth River near Walensee as 1:3.5, and in the Sarine River near Kallnach as

1:3.4. Fortier and Blaney estimated that one-fifth of the detrital load in the Colorado River where it passes Yuma, Ariz., is carried by traction. Sykes considered this ratio too small. It is probable that there is no fixed relation, and the writer has observed instances in which there was no apparent suspended load but a large tractional one, and cases have also been observed where the opposite seemed to be true. The suspended load is subject to equally wide variation. Streams from vegetable-covered areas may carry almost no visible suspended load for nearly every day in a year, whereas the ephemeral streams of arid and semiarid regions may have their channels in the form of dry coulees one hour, and the next they may be filled with water with such a load of inorganic sediments as to be almost a mudflow. W. Z. Miller gave the information that the suspended load in the San Juan River at Goodrich, Utah, at one time in 1929 was 40 per cent by volume. This was almost a mudflow. In general, the streams of arid, semiarid, and cultivated regions with flashy response to rainfall have the largest loads of suspended materials, and streams with least loads flow from plant-covered areas. The Rio Grande has records showing a suspended load as large as 10 per cent of the volume, whereas the Mississippi near its mouth has been stated to have an average load of 0.07 per cent and a maximum of 0.08 per cent of the volume. The maximum load is carried at times of flood waters. In the Fraser River of British Columbia this load was found to be 230 parts (dried at 180°C.) per million, whereas at times of low water this fell to as low as 10 parts per million.

According to the estimates of Russell the Mississippi River carries at least 2 million tons of sediment past the city of New Orleans to the lower delta each day. This is 83,000 tons per hour, and the total daily transportation is equal to nearly 80,000 gondola cars, each carrying 25 tons. A train carrying this burden through New Orleans would have a loaded car passing a given point in the city about every 11 seconds. This includes the sediments transported both by traction and suspension. Older figures place the figure at 400 million tons annually. Dole and Stabler estimate the annual suspended load of the streams of the United States at 513 million tons and in addition 270 million tons of dissolved matter (p. 83). Fortier and Blaney (p. 25) give the mean "silt load" annually transported by the Colorado

River past Yuma, Ariz., over a period of 18 years from 1903–1905 and 1911–1925 as 183,759,000 tons with a range from 91,348,000 tons in 1918 to 308,728,000 tons in 1905.

All phases of rock destruction produce colloidal matter. Many of the products of decomposition are colloids, and they are produced by abrasion, grinding, and impact. Silica, iron and aluminum hydroxides, hydrous aluminum silicates in the form of clay minerals, organic matter, and perhaps carbonates are the most common colloids transported. The silica, alumina, and ferric iron reported in the older analyses of water in large part are probably in the form of colloids and not in true solution.

It is probable that considerable quantities of colloids are flocculated below the junctions of any two streams in case the substances in solution are unlike, but flocculation is rather complete after stream waters reach the sea or salt lakes and the fresh waters mingle with salt. It does not necessarily follow that deposition takes place immediately following flocculation, as the low density of the flocculates permits them to be easily transported by waters of low competency, and the density of the salt waters works to the same end. Usually, however, there is so much macroscopic suspended matter that the flocculated colloids are rapidly carried down with these visible sediments and hence make no independent deposit. The precipitated colloids are in the form of gels and contain much water and hence serve as excellent cements for the materials with which they are associated. Hilgard records the presence of a 4.5-meter layer of jellylike material on the outer side of a bar near one of the mouths of the Mississippi River, and Russell found a jellylike deposit in the same area, which he assumed was largely composed of silica, and was stiff enough to stand alone. In southern Queensland, Australia, Barton has described bars of constant position at the entrance into the sea of the fresh waters of the Tallebudgera and other rivers. Maintenance of constant position was ascribed to precipitation of colloids that served as binders for the composing sands. If the loads of streams consist almost entirely of dissolved and colloidal matter, as would be likely of streams draining base-leveled areas covered with vegetation, such streams could easily make a deposit, composed of colloids of silica, iron hydroxide, manganese oxide, aluminum hydroxide, or some combination of these substances.

As soils of base-leveled areas should be rather thoroughly leached of soluble substances, few carbonates might be transported.

Many colloids are prevented from flocculating by the presence of protective or stabilizing colloids. These in some way prevent neutralization of the electric charges. Organic colloids seem to be particularly effective to this end.

Many substances carried by streams have the property of base or cation exchange. Among these are the montmorillonite-beidellite group of clay minerals, some of the organic colloids, some of the zeolites, and various silicates. These substances are present to a greater or less degree in most soils, and hence base exchange may take place where sediment-laden waters of different origins mingle. Following entrance of a tributary into a stream there should be changes in the sediments thus brought together and also changes in the dissolved contents of the waters until conditions of equilibrium have been attained. Results should be more conspicuous where streams enter the sea and fresh waters are mingled with salt. Elements readily subject to base exchange are calcium, magnesium, potassium, and sodium, named in order from highest to lowest replacing ability. Thus sodium clays, acted upon by waters carrying calcium, magnesium, or potassium, change to calcium, magnesium, or potassium clays until conditions of equilibrium have been reached. As the processes are reversible to a greater or less degree, each of the clays may change to any other. After the sediments are buried, the repeated entrance of new fluids may lead to further base exchange.

SOLUTION TRANSPORTATION

It is probable that every known substance is carried in solution. The solution load of any stream is determined by what is available in the region drained. Streams draining limestone regions of any relief are high in calcium and magnesium carbonates. Streams draining arid and semiarid regions are unusually high in dissolved materials, and, if a region has been arid in its past history so that evaporites are present in its geologic section, calcium sulphate and sodium chloride and other salts may be present in large quantities in the solution load. Streams draining areas underlain by igneous rocks ordinarily carry a small load of dissolved substances, and the waters tend to be soft. Streams draining plant-covered areas have the dissolved loads increased by the presence of the plants, as their decay produces carbon dioxide, which aids in placing carbonates in solution. Some part of the solution load of streams and standing bodies of water is acquired from the suspended and traction loads. The quantity is related to the degree of solubility of these loads and the extents of their transportation.

The ratio of the dissolved to the suspended load of streams draining regions of different rocks, climate, and plant growth shows interesting relations. Streams that drain regions covered with vegetation are likely to have a ratio between the dissolved and suspended matter that is much greater than unity, but the absolute quantity of sediment carried by either method of transportation is likely to be small, whereas streams draining dry, treeless areas have the ratio less than unity, although the loads of both dissolved and suspended matter tend to be large and many times the quantity of the corresponding loads in streams draining plant-covered areas. The ratio of the dissolved to the suspended matter in the Kennebec at Waterville, Maine, is 12, that of the Hudson, 6.7; the streams of the Middle Atlantic states north of Virginia have the two loads in the ratio of 2.9, but for the southern Atlantic states this ratio decreases to 0.82. The ratio for the East Gulf states is 0.93, and for the dryer states of the West Gulf, 0.71. The average ratio of the dissolved to the suspended sediments in the Mississippi below Minneapolis is 5.6, but at Minneapolis it is 25. The Missouri River, draining a much dryer and a nearly treeless region, has a ratio of 0.40. There is only a trace of suspended matter in the St. Lawrence River, which is a consequence of such matter having been left in lakes, but the dissolved matter ranges around 100 parts per million, thus giving a ratio that is much in excess of 100. The smallest ratio is that of the Rio Grande at El Paso, where the ratio is about 0.05, or there are about 14,140 parts per million of suspended matter to 700 parts per million of dissolved matter (Dole and Stabler).

These figures show that the suspended loads of rivers draining dry areas are larger than the dissolved loads, whereas, if the streams drain plant-covered areas, the dissolved loads exceed the suspended loads. Deposits made by rivers that drain dry areas might have sediments deposited from solution about their places of entry into the sea masked by the sediments deposited from suspension, whereas the deposits made by streams draining plant-covered areas might have the sediments deposited from solution masking those deposited from suspension. Under the latter condition such things as cherts, flints, and iron oxide readily become concentrated in the sediments, whereas in the former the silica and iron oxide are disseminated throughout sediments deposited from suspension.

Transportation and Deposition in Standing Bodies of Water

Transportation by traction and suspension in standing bodies of water, unlike that of streams where the transportation is downstream, may be back and forth over the same area for years, on the beach and back again offshore, one direction along the shore one hour or day, and in the opposite direction the next hour or day. Sediments may travel many miles and at the end be but a few yards from places of starting. Sediments from time to time escape from this seemingly aimless wandering and are carried out to deep water and permanently deposited.

Capacities and competencies are slightly increased in the waters of the sea and salt-water lakes because of the greater densities of these waters. Except near shore the currents in standing bodies of water have such low velocities that both capacity and competency are small. Adjacent to the shores there are extremely strong currents at times, moving onshore and offshore and also parallel to shore. These have great competency and capacity. Strong currents are often present in narrow passages between lands, and these currents oftentimes extend to great depths, as in Florida Strait, where the bottom is said to be scoured clean of fine sediments, and in the Canary Islands, where currents are said to extend to bottoms of depth of 2,000 meters.

Except in waters over shallow bottoms and near shores, the loads of standing bodies of water are small in terms of any unit of volume, but in view of the large quantity of water the total load almost staggers imagination. There is not a great deal of information respecting loads in the open ocean away from shallow water. Reade and Holland have expressed the opinion that the ocean "may hold, carry about and distribute such fine or microscopic sediments in quantities, if not commensurate with, second only to the matter held in solution." A sample collected by Murray from the North Atlantic showed a content of suspended solid matter of 1,601 tons per cubic mile; one from the Mediterranean 2,031 tons per cubic mile; and one from the Indian Ocean, 1,264 tons per cubic mile. A sample of sea water after being permitted to stand for 30 days still retained 625 tons of finely divided suspended matter per cubic mile. Assuming an average of 1,000 tons per cubic mile gives a total load of suspended matter in the deep waters of the sea of something over 300 billion tons. It would take the rivers of the United States over 100 years to carry these sediments to the sea. This material is mostly of colloidal dimensions, but some particles may be fairly large and be transported because of being enmeshed in organic jellylike substances like that described by Raymond and Stetson from Massachusetts Bay and by Russell from the Mississippi Delta. Long transportation favors solution.

Waves of translation and currents often carry very large loads over shallow bottoms. Studies made on the coast of Holland by Vervey found that at high tide in a calm sea 1 cubic meter of surface water contained 109 grams of sand and 1,303 grams of clay and silt. At ebb tide with a calm sea 1 cubic meter contained 304 grams of sand and 1,094 grams of clay and silt. At a distance of 1 meter from the bottom at high tide the quantity in suspension in 1 cubic meter of water was 1,094 grams of sand and 1,861 grams of clay and silt; at ebb tide the same distance above the bottom the suspended matter consisted of 1,062 grams of sand and 2,980 grams of clay and silt. These are large loads, and, if representative of the waters bordering all shores of the land, they emphasize the tremendous transportation of suspended matter in shallow sea and lake waters. Few of the sediments transported in very shallow waters attain permanent deposition there. They are more or less constantly in motion, but from time to time some escape the apparently aimless travel and are carried to deeper bottoms and permanent deposition.

Waves roll on shores until all energy is exhausted and the landward motion of the water is brought to a halt. These waves bring with them particles in terms of their competency and capacity. The pull of gravity draws the water back, and under most conditions this pull is less than the force that drives waves onshore. The result is that the returning waters have neither the capacity nor the competency to carry back all that was brought. Something may be left on the beach and it is that which the returning waters are unable to carry back. The incoming wave is checked progressively as it passes up the shore, so that it begins to deposit its tractional load with the first checking of bottom velocity, and when the water has attained its highest position and is momentarily stationary all except the very finest sediments are deposited. The substances last deposited are often organic, so that the position of highest advance is commonly marked by a line of organic materials. The returning waters gather momentum as they flow down the beach, so that progressively larger materials are picked up, but the coarsest material carried to each level of a wave's progress upward may be expected to be left. There is another factor that still further weakens the competency and capacity of the returning waters. As the waves roll upward on a beach some water is absorbed by the materials of the shore, and the water so absorbed does not return as the waves move back, but seeps out in the form of small rills between each onrush and backflow. The waters in the rills carry very little on their return.

This onrush and backflow of waves is responsible for much abrasion, grinding, and impact, and large quantities of finely divided sediments are produced, which in course of time are carried to deeper waters. A count of the number of times per minute small waves roll onshore in San Francisco Bay near Palo Alto, Calif., gave an average for a number of counts of 22, and on the shore of Planting Ground Lake in northern Wisconsin the average was between 30 and 40.

The general fact of waves moving both coarse and fine material to the beach and returning seaward only the finer of that brought is a matter of great importance in the deposits of large bodies of water. As a sea with stationary sea level progresses inland through erosion of shores and bottoms it cuts a bottom surface that should not have any sediments upon it other than pieces dislodged from the floor or shore that are too large to be moved. The coarse sediments are brought to the beach and remain there until abrasion, impact, and grinding reduce them to such dimensions that the returning waters can carry them away and give some of them permanent deposition in deeper waters beyond the bottom cut by erosion. Should sea level rise, deposition in course of time would invade the outer margin of the surface eroded.

Most waves strike a shore obliquely, and thus the particles brought to a shore are at the same time moved parallel to it. The returning waters with their burden of sediments move directly downshore to be hurled upward again by the next wave. The result is that suspended and tractional sediments move in zigzag fashion parallel to a shore until an indentation is reached where they may be transported to a bayhead whence outgoing currents in the bay may take the finest to deep water and leave only the coarsest. However, instead of entering the bay, the conditions may be such as to transport the sediments into the water across its entrance to form first a submerged ridge, then a spit, and perhaps in course of time a bar.

Shepard, Emery and LaFond have called attention to the transporting ability of *rip currents* in moving sediments from shores to deeper waters. These currents move outward from shores in the form of lanes somewhat normal to the beach. The rate of flow is very inconstant, and the water is generally strongly agitated and may contain much sediment. They suggest that the rip currents may have been identified as undertow.

SHORE PROFILES DEVELOPED BY EROSION AND TRACTION AND SUSPENSION TRANSPORTATION

Erosion by waves and currents on shores and bottoms and traction and suspension deposition in deeper waters seaward of areas of bottom erosion ultimately develop a surface over which outgoing currents when loaded with sediments to capacity make no deposits and produce no erosion. When a bottom surface has attained this level it is said to have a *profile of equilibrium*, which means that it is a surface adapted to the conditions existing at the place. This surface is a *temporary base level of erosion* where it has been made by erosion and a *temporary base level of deposition* where it has been built by deposition. If a bottom has any parts above the level of a profile of equilibrium, they are eroded until that level is attained; and if there are places that are lower than that level, deposition takes place until the bottom has been built thereto. A temporary base level of erosion may be just below low-tide level.

Any profile of equilibrium persists only so long as the conditions to which it is due exist. Any erosion on the shoreward side, or any deposition on the seaward side makes necessary an adjustment of the profile to a different level. A time of weak outgoing currents and a large supply of sediments lead to deposition on a part of the bottom previously at the level of the profile of equilibrium, whereas strong waves and currents with a small supply of sediments lead to erosion on the bottom. The result is that there is much temporary deposition and erosion on bottoms that have attained the approximate level of a profile of equilibrium. If conditions develop whereby outgoing currents cross a bottom unloaded, this bottom would be eroded and in course of time cut to *wave base* or the *permanent base level of erosion.* The sediments thus acquired are transported outward to the bottom of the water body, which, so far as possible, would be built to the level of the *permanent base level of deposition.* The maximum depths of base levels of erosion and deposition naturally vary with conditions, but both are generally above the lower boundary of the neritic life zone. The base level of erosion reaches best development in the old-age stage of the cycles of shore and land erosion. When the base level of erosion has been reached, all deposits of bars, spits, bayhead beaches, and deltas have been removed, so far as they are above this level (Fig. 25).

After a bottom has attained the level of a profile of equilibrium and as it progresses to the level of the permanent base level of erosion, no permanent deposits are made on the bottom, but temporary deposits may be, and the bottom is a base for continuous shifting of sediments seaward. The bottom throughout the entire sequence of events is an eroded one, and the time involved in its erosion is not recorded at the place. The top sediments on this bottom are those that the outgoing currents could not carry away, and these tend to be coarser than the sediments immediately below.

If conditions develop to bring about rise of sea level, the bottom may receive sediments and be built to a new level—a new permanent base level of deposition. The new deposits are separated from the old by a stratigraphic break, and they are likely to be finer grained than those upon which they rest. As the underlying sediments may not be cemented, some of the new deposits may sift into them to place sediments and microorganisms of a later time in the interstices of sediments of an earlier time. The microorganisms of the new deposits would indicate one time relationship, the macroorganisms of the older

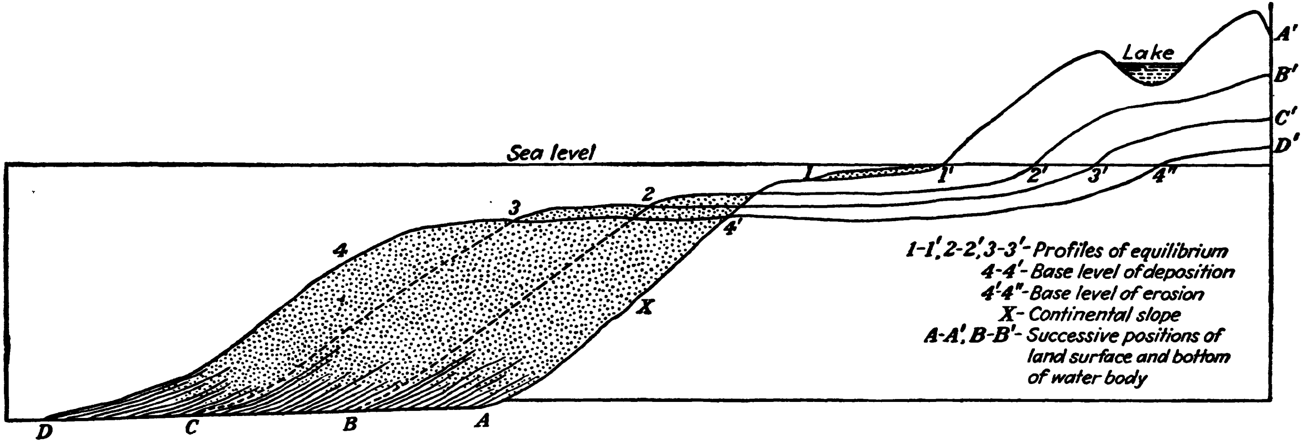
Sea level
Lake
A'
B'
C'
D'
1'
2'
3'
4''
I
1
2
3
4
4'
X
A
B
C
D
1-1', 2-2', 3-3'- Profiles of equilibrium
4-4'- Base level of deposition
4'4''- Base level of erosion
X- Continental slope
A-A', B-B'- Successive positions of land surface and bottom of water body

Fig. 25. Diagram showing progressive positions of a profile of equilibrium with stationary sea level and during reduction of a land area. The base level of deposition is indicated by 4–4′, the base level of erosion by 4′–4″. The final profile of equilibrium coincides with the position of the base level of erosion and the base level of deposition.

deposits an older time relationship. But both are in the same lithic unit.

Slopes of profiles of equilibrium vary with the conditions to which they are due. They tend to be more steeply inclined adjacent to the shore in large bodies of water in which waves and currents are strong, and more gentle in small bodies with weak waves and currents.

What has been said relative to base levels of erosion and deposition and profiles of equilibrium applies to the open sea where deep basins which cannot be filled with sediments lie outward from the shallow waters over the continental shelves. The terms have application during the early and probably middle histories of such interior seas as Hudson Bay and the Baltic Sea and those bodies of water that lay over the Mississippi Valley and the Appalachian geosyncline during the Paleozoic, but the terms cease to have application when the deeper parts of such basins become filled to the extent that sediments cannot be shifted from the shallower sites of deposition. Thence onward—unless there is a sinking of the bottom—the sediments would be spread widely and might fill the basin to some level above that of the sea to which the body was tributary. The site of an interior sea would then become a vast swamp as seems to have been the final stage of the Appalachian geosyncline during the Pennsylvanian period.

DISTRIBUTION OF SEDIMENTS IN STANDING BODIES OF WATER

It seems probable that deep-water sediments far from land are deposited in sheets of essentially uniform thickness and character over great areas. Bottoms that have attained the level of the profile of equilibrium and the base levels of erosion and deposition receive no sediments for permanent deposition. Between these two extremes are other bottoms upon which the sediments deposited are responses in texture and quantity to changes in competency and capacity. As these change from day to day and from hour to hour and frequently from moment to moment at any given place and are rarely the same from place to place in any direction, it results that the sediments deposited in these waters pass laterally into other sediments of other character and thickness or pass to bottoms upon which no deposits are made. Beaches also receive deposits. As previously shown, the beaches are composed of sediments that the returning waters could not carry away; and if the beach sediments came from the sea, they consist of the coarsest sediments brought by the waves. It is obvious that the materials of the beach are functions of the waves that exist there, and thus beach sediments may range from clays to boulders; and as the competencies of the incoming and outgoing waters vary from time to time and from place to place, it results that some beaches are composed of like materials for long distances whereas the sediments of other beaches change every few feet. However, many beaches derive materials from the lands. This does not alter the situation except that it introduces another factor to increase the variations. A characteristic of beach sediments is shingling. This requires particles several inches in diameter and with two long dimensions and one small. These tend to incline toward the water and overlap each other from the upper to the lower part of the beach like shingles laid from the comb to the eaves of a roof. This arises from the fact that the incoming waves are stronger than the outgoing currents. If the outgoing current were the stronger the overlap would be in the opposite direction (Fig. 24). This is locally the case.

There is a general impression that clastic sediments decrease in coarseness from the shore outward, and this concept in its broader outlines is probably correct on some coasts, but there are many exceptions, as is shown on Shepards's maps of the distributions of sediments on sea bottoms.

Trask and Revelle have shown that the physiography of a bottom is a factor of first importance in the distribution of clastic sediments. In the region of the Channel Islands of California the coarser sediments are on the high parts of the bottom and the finer are in the basins (Fig. 26). The reason for this distribution is the selective action of waves and currents in moving fine materials to the extent of competency and leaving the coarse materials because of incompetency to transport. The physiography of a shore and the impingement of the tides in channels are also factors. In Bering Sea and Davis Strait in depths of 900 or more meters and on bottoms many miles from land the sediments are sands, whereas in shallower waters they are silts and clays. In places in San Francisco Bay muds are deposited inshore, sands farther out, and gravels in the deeper waters (Louderback). This arrangement is doubtless due to the occurrence of stronger currents in the deep than the shallow waters.

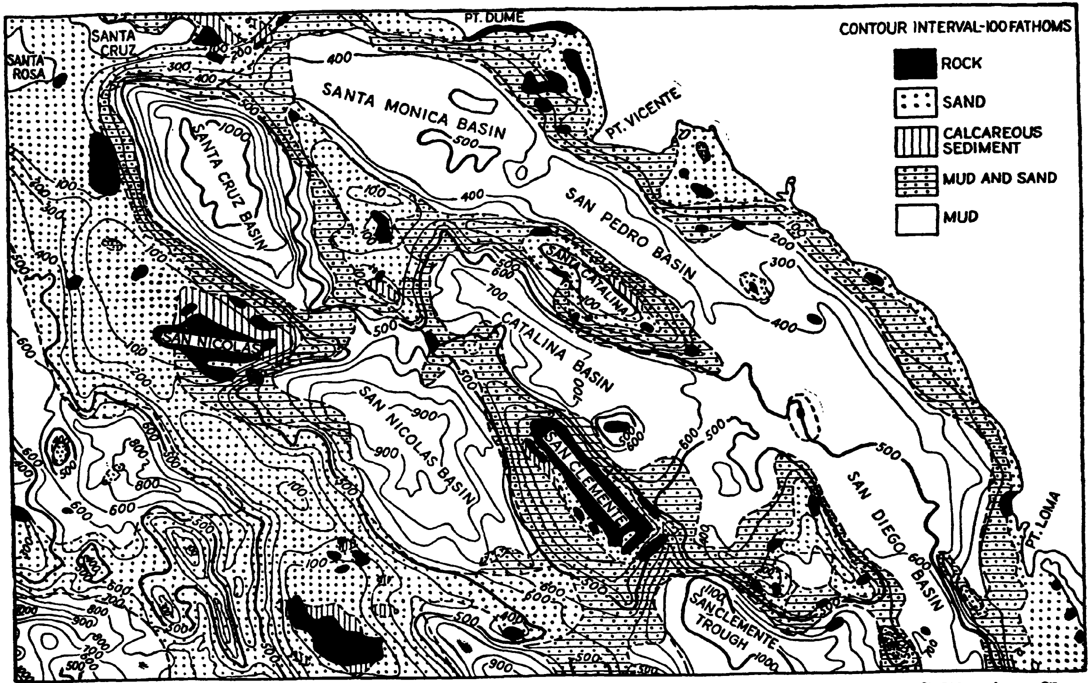

FIG. 26. Distribution of types of sedimentary deposits in the offshore region of southern California. Close dependence of sedimentary processes upon topography of the bottom is shown. The white areas representing muddy bottom are nearly uniform depressions on the sea bottom. The elevations surrounding the depressions are covered with more or less coarse sediments, or are bare rocks without sedimentary cover. (*Redrawn from a photograph; courtesy, Roger Revelle, Scripps Institute of Oceanography, La Jolla, California.*)

Organisms are not generally distributed as to kinds and numbers over the bottoms of any water body, and scavengers are present in more or less direct ratio to the other organisms upon whose dead bodies they feed. The results are that organic sediments are ground up and evidence of organic origin may be lost (Wepfer, 1916). However, sediments of the sea bottoms of the past are often filled with organic remains in excellent preservation. In some cases this may be due to such lush growth as to be beyond the capacity of the scavengers to destroy the shells. This evidently was the case on the *Pentamerus Banks* common in Silurian strata. A large influx of muds or sands, however, might completely bury the bottom organisms: herbivores, predators, and scavengers. Vagrant forms in some instances might extricate themselves if not buried too deeply, but fixed forms such as coelenterates, brachiopods, bryozoans and others could not do this, and they would remain in the sediments in positions of life to make a wonderfully fossiliferous layer that would be composed of clays, silts, or sands between calcareous layers composed of the comminuted remains of organisms. The calcareous layers record the normal conditions of slow deposition, the fossiliferous layers the occasional incidents bringing about rapid deposition. This explanation may account for the richly fossiliferous layers often seen in geologic sections where more or less unfossiliferous limestones are separated by layers of shale filled with fossils. Thin sections commonly show that the limestones are composed of the comminuted remains of shells. Excellent shells may be found adhering to the surfaces of the limestone beds. The shells seem to be more common on upper surfaces than lower, probably owing to the fact that upper surfaces were buried by the sediments (Wepfer).

RATES OF DEPOSITION IN STANDING WATER

Little is known respecting rates of deposition of sediments in standing bodies of water, and much that has been published is of little value. Attempts to arrive at average rates of deposition have been made along three lines: (*a*) The assumed rate of erosion of any given region has been compared with the area of the basin receiving sediments. This may have some doubtful applicability to existing areas, but it has no application to sediments of the past where neither the extent of the area undergoing erosion, the rate of erosion, nor the area of the basin of deposition is known. Any figures obtained in this way are little more than very poor guesses (Walcott). (*b*) The thickness of the column of sediments is divided by the length of geologic time. This gives a figure representing the average rate of deposition for the geologic time represented by the column, but probably does not approximate correctness by an extremely wide margin, as no consideration can readily be given to the times of no deposition represented in every geologic section of any thickness. (*c*) The third method is based on observation, and any conclusions based on the data can be considered to apply only to the places where the data were collected and not elsewhere. Data are few. It has been determined that on some coral reefs building upward is at the rate of 1.5 mm. per year or 30 cm. in 190 years. Vaughan estimated the rate of growth of *Orbicella annularis*, the principal builder of reefs in the West Indies, as 5 to 7 mm. per annum, which at the average rate of 6 mm. per year would form a reef 45 meters thick in 7,620 years. *Acropora palmata* grows more rapidly and could form a reef 45 meters thick in 1,800 years. Gardiner estimated that in the Pacific Ocean a reef 45 meters thick might form in 1,000 years. The rate of growth in the fringing reef at Oahu has been estimated as approximately 30 cm. in 300 years. It is probable that lime sediments are not deposited under many conditions as rapidly as they form about coral reefs.

Sediments deposited from suspension and traction may accumulate with great rapidity, and rates of deposition of 0.33 meter in one day are known in places on the Mississippi Delta. A rate of 6 meters per year is known at the mouth of the Fraser River in British Columbia (Johnston).

It is probable that there are no places, or at least very few, where rates of deposition are so slow as beneath the waters of the deep sea. Estimates of the rate of deposition over the deep bottoms range from 1 cm. in about 500 years to the same thickness in about 3,000 years (see page 138). According to Schott (1939) the rate of deposition of Globigerina ooze in the southern Indian Ocean since the end of the Pleistocene ranged from 1/4 to 1 cm. per 1,000 years and of diatom ooze from 1/4 to 7/10 cm. per 1,000 years. In the equatorial Atlantic Ocean since the end of the Pleistocene, Globigerina ooze has been deposited at the rate of 0.53 to 2.13 cm. per 1,000 years. Blue mud and red clay in the same region had average rates of deposition of 1.78 and 0.86 cm. per 1,000 years, respectively. As the sediments do not seem to have been water free, the actual rates of deposition in terms of solid sediments must have been much slower.

With the wide ranges in estimated rates of deposition it seems a vain hope to arrive at an average one, and if such were obtained, it would have little or no value. On the Island of Timor in the East Indies the Upper Triassic is only 2.1 meters thick, but this thickness represents millions of years of time. The Tertiary of the Ventura Quadrangle of California is 9,000 meters thick, but this thickness probably represents no longer time than the 2.1 meters thickness of the Upper Triassic of Timor.

Various assumptions have been made as to the relative time values of limestones, mudstones, and sandstones. Most of these assumptions have little value. Ulrich assumed that 1 ft. of limestone represents the same duration of time as 7 ft. of mudstones and sandstones. Smith placed the ratio for the same kinds of rocks as 1 to 10. This fixing of a ratio totally disregards the fact that the rate of deposition is dependent upon supply and the positions of the sites of deposition with respect to base levels of deposition or profiles of equilibrium. *A few centimeters of cross-laminated and current-ripple-marked sandstone may be deposited on a surface near the level of a local or final base level of deposition, while at the same time on a bottom considerably below such levels many meters of limestone may be deposited. These ratios are valueless.*

THICKNESS OF ACCUMULATION

The thickness that may be accumulated and the time required to accumulate a thickness of sediment depend upon the two factors of the quantity of sediments brought to the sites of deposition and the position of these sites with relation to temporary or permanent base levels of deposition. If a large supply is brought and cannot be removed, sediments accumulate rapidly until built to the base level of deposition; and, if the surface of deposition settles as the level of the base level of deposition is approached, an extremely great thickness at a rapid rate would be possible. If it does not settle, deposition ceases at the place when the base level of deposition is attained. This may perhaps be illustrated for a special case by reference to coral reefs of the

Pacific to which Gardiner ascribed a rate of deposition of 0.3 meter in 1,000 years. The surface of the water in this case is the base level of deposition. If only a single foot separates the top of a coral reef from the surface of the water at a certain place and time, the two would coincide 1,000 years later if sea level remained stationary, and they would still coincide 50,000 years later; but elsewhere where 25 meters originally intervened between the top of the reef and the surface of the water there might be 25 meters of rock. This consideration indicates the absurdity of fixing a comparative rate of deposition for the various varieties of sediments unless all conditions of deposition are fully known.

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ATMOSPHERIC TRANSPORTATION AND DEPOSITION

Atmospheric transportation and deposition differ from that done by other agencies in that the sediments may be deposited at elevations that are considerably higher than sources. The transportation is like that done in the sea in that the sediments move in various directions, but differs in that, whereas in the sea the sediments may ultimately move to the deep basins, the atmospheric sediments have no definite goal. Most of the load is acquired through the direct work of the atmosphere. An unknown small quantity enters the atmosphere from outer space. Volcanic eruptions at irregular intervals provide "ash" and dust, and human industry pours vast volumes of dust from chimneys. In general, the atmosphere is at a disadvantage in obtaining a load in that its lowest velocities are adjacent to the ground. Measurements made on the Eiffel Tower, about 300 meters high, indicate velocities four times greater than on an adjacent tower 21 meters high. Much aid is given the atmosphere in acquiring a load by the great herds of animals that are not uncommon on dry prairie lands, and human activities on roads and in cultivated fields work to the same end.

The currents of the atmosphere are extremely complex, particularly adjacent to the ground where the irregularities develop turbulence, so that the currents move in various directions: up, down, laterally, and in whirls. In some cases these currents aid each other and in others they bring about neutralization. They vary in velocity and direction from moment to moment as shown in Fig. 27. It is this turbulence in the atmosphere that permits it to lift materials from the ground, and this lifting ability perhaps attains its maximum effectiveness in the many whirlwinds characteristic of dry areas. Jutson states that these "dust whirls," as he terms them, are important in subarid western Australia, and he noted the occurrence of one that rose from 3 to 6 miles. They are locally termed "willy-willies." Surfaces with considerable irregularities bring their high parts into the region of high velocities, and the irregularities produce diversity of currents and great lifting ability. High places,

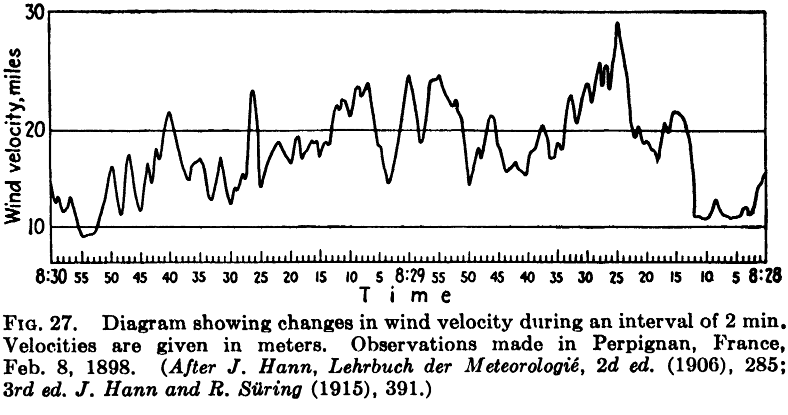

FIG. 27. Diagram showing changes in wind velocity during an interval of 2 min. Velocities are given in meters. Observations made in Perpignan, France, Feb. 8, 1898. (*After J. Hann, Lehrbuch der Meteorologié, 2d ed.* (1906), 285; *3rd ed. J. Hann and R. Süring* (1915), 391.)

accordingly, lose all fine materials almost as soon as these are detached from parent rocks.

Surfaces without vegetable cover yield the greatest part of the load that is naturally obtained by the atmosphere. Little is obtained from plant-covered surfaces. Uncovered surfaces are those of deserts, mountains, sea and lake shores, flood plains, surfaces and outwash plains of glaciers, and, since the development of large-scale agriculture, cultivated fields. Dry lakes of deserts are among the common natural areas in producing dust, in that the sediments in some are broken into fine particles through crystallization of contained salts.

The atmosphere transports sediments by traction and suspension. Sands and occasionally larger particles are transported by traction, and dusts and, in cases of high velocity, sands and larger particles are carried in suspension. Traction transportation may be individual or collective. In the former the sands travel as single individuals, of which there may be many in movement at the same time. Collective movement takes place in the structures which range from dunes to wind ripple marks. The entire body travels through movement from time to time of some of the composing particles. The small structures are comparable to the "dunes" formed in aqueous traction and are due to the same causes, whereas the larger structures primarily may owe their origin to some obstruction on the surface. The sand grains are rolled up the windward sides of dunes through the push of the wind and roll down the leeward side under the influence of gravity to produce cross-lamination on that side with inclination in a leeward direction. Each lamination or bedding unit is due to wind of a certain velocity, and the sand grains differ more or less from those in the laminations above and below in size of particles but not in composition. The laminations on the leeward side are inclined at angles that approximate the angles of repose for the composing particles with some modification due to the winds that brought the sands to the crest and to reverse currents up the leeward slope, which theoretically are present. The angles of repose are expressed in the formula of Thoulet (1887):

$$AR \alpha \frac{dm(dm - dl)c \times r}{dl \times s \times ag \times d}$$

in which dm = the density of the material.
dl = the density of the medium.
d = the dimension of the particle.
s = the sphericity.
ag = the agitation.
c = the cohesion.
r = the rate of deposition.

According to Thoulet the inclination never exceeds 41 deg. This inclination is not known to have been seen in nature, but Passarge gives inclinations of 30 to 33 deg., and Cressey states that the inclinations in the Indiana dunes on Lake Michigan are never steeper than 32 deg. The high sphericity of some sands tends to lower the inclination. The larger particles are in the bottom part of a dune deposit, as they roll farther down the leeward slope than the smaller and are less affected by the return current up that slope. Inclinations tend to be higher at the top of a dune.

Inclinations tend to be higher in air than in water due to the low density of the medium. General appearances are the same. Planes bounding cross-laminated units made by wind deposition tend to be at various angles, owing to the varying directions and velocities of the winds, whereas in water cross-lamination they tend to be parallel to the horizontal. Thus, wind cross-laminated units are wedge-shaped if observed in two dimensions, whereas this is not likely in aqueous cross-lamination.

The sizes of particles transported by winds vary with velocity. Udden made observations of distances of transportation in moderately strong winds and found that silts and clays are readily transported in suspension for hundreds of miles, but that sands and larger particles soon settle to the ground, where, however, they may be readily transported by traction. It is known that particles of silt and clay have been transported thousands of miles.

The competency of atmospheric currents is much less than that of water because of lower density. There is some compensation, however, in the generally higher velocities of winds and the greater turbulence produced by irregularities of the surface. Particles moved by traction are rarely larger than sands, but Pumpelly relates seeing particles 5 cm. in diameter blown by a storm in Turkestan, and Blackwelder states that a wind in the

Rogers Playa of the Mojave Desert of California with a velocity of 40 m.p.h. at a height of 1.5 meters above the ground rolled pebbles to 10 mm. in diameter and threw particles 2 to 3 mm. in diameter into a touring automobile. Wind ripple mark with wave length of 2.4 to 3 meters and about 25 cm. high was seen in this playa with the composing particles 1 to 4 mm. in diameter, and, in Coyote Playa in the same desert, wind ripple mark was seen in which the particles ranged from 5 to 23 mm. in largest and 4 to 10 mm. in shortest diameter.

Various experiments to determine competency have been made, among which are those of Udden, Sokolow, and Thoulet (1908). Beadnell has stated that sands begin to move when the velocity of the wind attains 13 m.p.h., and the air is visibly charged with sands at 23 m.p.h. Experiments with winds of known velocity gave Thoulet the results shown in Table 16.

The figures indicate that competency is in about direct ratio to velocity. Twelve meters per second is between 26 and 27 m.p.h., and this is competent to transport sand grains in suspension. It is obvious that winds with velocities of 100 or more m.p.h. can readily transport granules and pebbles.

TABLE 16. COMPETENCY AND WIND VELOCITY

Velocity, M. per Sec.	Average Diameters of Particles, Mm.
0.50	0.04
1.00	0.08
2.00	0.16
4.00	0.33
8.00	0.65
12.00	0.97

The capacity of the atmosphere per unit of volume is far less than that of water, owing to the fact that the density is $\frac{1}{813}$ that of water. However, as capacity is so largely dependent upon velocity, the fact that the average velocity of winds is much greater than that of water reduces the difference very materially, and at some velocity the capacity of the atmosphere per unit volume may be greater than that of ordinary velocities of water. It is generally stated that capacity varies as the sixth power of the velocity. It is not known that such is the case, but if so, wind velocity need become from three to four times that of water for the capacity per unit volume to become equal. The atmosphere over the lands has far greater capacity than all the land waters, and Udden (1894) stated that the capacity of the atmosphere over the Mississippi Valley is 1,000 times the capacity of the water on the surface over the same area.

Much of the atmospheric capacity is unused, but in spite of this fact the quantity of sediments carried is extremely great, as the dust clouds of the past several years demonstrate. Counting of dust particles (Aitkin) in the atmosphere has shown that under ordinary conditions the number of particles per cubic centimeters is large, and even atmosphere that seems clear may be seen to be filled with dust particles if a room is darkened and a beam of light is passed through the air. Udden (1896) estimated that 850 million tons of dust are carried an average distance of 1,440 miles by the atmosphere over the Mississippi Valley each year. The great dust storm of May, 1935, suspended dust to the extent of 101 tons over each square mile in the region of Washington, D.C., with only the height to 1 mile considered (Hand). This dust came from the western plains. Lugn estimated on the authority of Ver Wiebe that in the great dust storm of Mar. 20, 1935, there were 5 million tons of dust suspended in the atmosphere to the height of 1 mile over the 30 square miles of the city of Wichita, Kan., and, as dust extended upward to 12,000 ft., the total quantity in suspension must have been greater than the figure given. Lugn estimated that the dust deposited over the area around Lincoln, Neb., amounted to about 40 carloads per square mile. Free states that Europe has received 5½ in. of dust from the Sahara since the beginning of the period of record. According to Petrie deflation has removed more than 2.4 meters of thickness from the Nile Delta in 2,600 years, and Blackwelder has shown that more than 4 meters have been removed by deflation from the bottom of Danby, a dry lake in the Mojave Desert. Some students have urged that many of the depressions of the Great Basin Region are due to deflation, and deflation basins in the Desert of Gobi that are 270 meters to 30 or more miles long and 15 to 120 meters deep eloquently emphasize the quantities the atmosphere actually may transport, even if the actual load per unit volume is small (Berkey and Morris).

The materials transported from any place depend upon what is available. The suspended load is composed of organic as well as inorganic materials, the latter consisting of all kinds of minerals, but mostly clay minerals, carbonates, quartz, and feldspar, and the former of plant tissue, diatoms, shells of protozoa, and spores. The minerals in many cases are fresh and unoxidized. Sands transported by traction are most commonly composed of quartz, but calcite, clay (Coffey), gypsum (Herrick, Talmage), limestone, chert, feldspar, and garnet sands are not uncommon. Some are composed of the tests of Protozoa, as on Dog Bay in Ireland, and fragments of shells, as on Bermuda and elsewhere.

Absence of the cushioning effect present in water impact and abrasion permits particles transported by the atmosphere to attain a higher degree of rounding to smaller dimensions than can be done in water. Most of this is done in traction transportation, but some may be accomplished on suspended particles, as the turbulence of the lower atmosphere with its maze of crisscrossing currents doubtless produces many collisions with mutual abrasion and impact of particles. In traction transportation, particles are in direct contact with each other and the underlying surface, and the larger particles may become rounded to a relatively high degree of sphericity. Ziegler has stated that if particles less than 0.75 mm. in diameter are rounded, it should be postulated that wind is responsible. Galloway's experiments led to the view that water may round particles to 0.05 mm., but that wind may be effective to 0.03 mm. It is generally believed that if particles are rounded below dimensions of 0.1 mm. and quantities are large, the rounding was done by aeolian agencies. If quantities are small, other agencies may be responsible. Reference to the formula given on page 216 shows that distance traveled is important in development of rounding, and as this in wind transportation is ordinarily much greater than in transportation by water, a greater degree of rounding is favored. The high degree of sphericity present in the sands of some dunes may be due to selective sorting, as suggested by MacCarthy and Huddle.

The repeated impact of particles on each other without cushioning effect produces minute chattermarks, fracture pits, and scratches, giving the ground-glass or *mat* surface on the particles, which are usually described as *frosted*. As there is no solution, nothing of this is removed as may be done in water. The presence of a mat surface on particles of sand or smaller dimension is considered to indicate that nonaqueous agencies, probably aeolian, have been concerned in the production of the rounding. MacCarthy's studies led to the view that the sands of the seashore dunes are more rounded than those of desert dunes.

It must not be assumed that deposits composed of highly rounded and frosted particles of small dimension are necessarily of aeolian deposition. Aeolian agencies may have been responsible for the detailed characters of the sands, but the winds may have given them to water, which thus became responsible for deposition. On the other hand, the deposits made by aeolian agencies are very often composed of angular particles. Transportation may have been short. Many sea- and lake-shore dunes are composed of particles that have few or no characters for which wind may be held responsible.

The surfaces over which aeolian agencies transport sands are modified by the movements. If the surface is based on solid objects, the sands cut furrows or grooves along lines of weakness. Where these have been cut on clays and silts they are known as *yardangs*. Similar grooves on hard snow and ice are termed *zastruga*. Hard rocks are etched, and, as movement is most intense close to the ground, the bases of high places are cut away to form the mushroom forms characteristic of desert regions. The cutting done at the bases of cliffs may form caves, of which the walls are due to abrasion and impact and not to solution.

Pebbles, cobbles, and boulders of more or less homogeneous composition in the paths of wind-driven sands have facets cut upon them to form the well-known *ventifacts* (Bryan) perhaps better known as *dreikanter*, *einkanter*, etc. Some facets may have pitted surfaces due to lesser resistance to abrasion and impact. This characteristic of pitted facets distinguishes ventifact facets from those cut by ice.

The so-called desert varnish has its development favored by aeolian abrasion and impact. The feature, however, is not confined to deserts, and something very similar forms on wave-beaten seacoasts and adjacent shallow bottoms through deposition from water of maganese oxide on the surfaces of materials of the beach and bottom. Desert varnish on the land seems to be due to capillary action bringing materials from the interior of a rock to the surface, which may then be polished by wind-transported particles. The shiny coating is composed either of iron or manganese oxide. It has been postulated that lichens by means of acid secretions have been largely responsible for transference of the iron and manganese oxides to the surface (Laudermilk).

Sediments deposited by the atmosphere are ordinarily better sorted than those deposited by water. The suspended load is usually a small part of the capacity, and the slowing of the velocity is so gradual that particles in the upper atmosphere settle in terms of the competency and capacity over a long distance and rarely bring down smaller particles with them in settling. If settling is forced by rain or snow, all grades are brought down together, but as a rule the deposits made in these ways do not persist. They are usually carried away by water. However, extreme perfection of sorting rarely obtains, since the particles as they approach the surface enter the maze of conflicting currents found there, in which competencies vary from nothing to the maximum. A hundred cubic meters of the lower atmosphere may contain particles of which the largest according to Udden (1898) may be 100 times the smallest. However, all particles are small, and departures from excellent sorting are not apparent to the eye, but they become obvious under a microscope.

Sediments transported by traction move often, and, as velocities greatly vary from time to time, the result is that the sands left at any place are repeatedly winnowed, all particles within the competency of the winds being removed, so that what remains represents differences of competency of the agents transporting to and from the place. The usual result is that the sands are excellently sorted if only the deposits of a short period of time are taken. Udden's (1898) observations show that the sands of dunes have a dimension range from 1/16 to 1 mm. with over 80 per cent between the dimensions of 1/8 and 1/2 mm.

Study of many analyses of dune sands by Udden indicate in the grades used by him that the grade second in quantity is on the coarser side of the maximum, which he explained as due to the ease by which coarser sands are rolled by traction, whereas the finer sands are lifted by suspension and removed. There is thus a relatively sharp break in grain quantity at that dimension where suspension competency begins. Dake stated the principle as the total quantity coarser than the maximum grade is larger than the total finer than that grade. Too close adherence to this generalization should not be made in interpretation of ancient sands, as water is known to deposit sands of similar sorting.

All noncolloidal sediments carried in suspension by wind are deposited when velocity and turbulence fall to zero. The small noncolloidal particles fall in accordance with the Stokes formula. However, there are so few times that the atmosphere above any place is absolutely calm and there are so many small diverse currents near the ground that settling does not proceed regularly. Rain and snow are very effective in forcing settling, but the particles then become water transported. Colloidal particles remain indefinitely in suspension unless washed out by rain or snow. Physiographic features are frequently responsible for checking velocity, and Willis noted that in China hollows in hills or high ridges transverse to winds are responsible for deposition of dust. Tractionally transported particles are deposited as velocity decreases, and all movement ceases with calm.

Suspension transportation carries small particles over the entire earth, and one may agree with Salisbury that every square mile of the earth's surface has acquired wind-borne sediments from every other square mile.

Sediments are washed from the atmosphere every time rain or snow falls. The coarser parts of the suspended loads are deposited leeward to places of acquirement, often hundreds and even thousands of miles distant. The great dust storms of 1934 and 1935 acquired dusts from the so-called dust bowl of southwestern Kansas and adjacent parts of New Mexico, Colorado, and Oklahoma and carried some of these beyond the eastern shore line of the United States, and, even as far distant from the source as the region of Washington, D.C., the atmosphere to a height of 1 mile contained 101 tons of dust over each square mile of surface (Hand). Volcanic dust is known to have been transported thousands of miles, and it is estimated that the dust produced by the explosion of Krakatao in 1883 encircled the earth. During the Pleistocene the high-pressure areas over the ice sheets produced south-blowing winds. These carried dust from the ice sheets and outwash fans, and this settled over more moist areas to the south.

The colors of the dusts transported by the atmosphere have a considerable range. Some are gray, others are yellow, and some dusts of southwestern United States are red. Red dusts are also known from the Sahara. The colors are due to the sources from which the dusts are derived.

The tractional load of the atmosphere generally moves in the form of dunes either of the barkhan, seif, or some complex form. Dunes do not travel so far or so fast as the dusts in suspension, and ordinarily the distances are of the order of magnitude of a few miles and at most a few tens of miles. The shortest distances are traveled by the dunes of humid regions, where the sands are derived from sea, lake, and river shores. These travel inland and ordinarily do not move more than a mile or so from sources, as they are soon stopped by vegetation. There may have been dunes about some margins of the Pleistocene and older glaciers for which the sands were obtained from outwash fans. Dunes of arid regions commonly margin low places and leeward borders and windward slopes of elevations. There is a great dune area east of the Caspian Sea. Sands, originally of stream deposition on the floodplain of the Jaxartes (Syr Darya) River, form into dunes that move southward across the Kizil Kum Desert into the Oxus (Amu Darya) River, which in turn deposits them on its floodplain, where they again form into dunes, which migrate across the Kara Kum Desert to the shores of the Caspian Sea. The average rate of movement of the dunes of Kizil Kum Desert is placed at 6 meters per day with a maximum of about 20 meters per day.

Rates of deposition of sediments carried in suspension in the atmosphere under ordinary circumstances are very small. The possible fall of the 101 tons suspended over each square mile of the Washington, D.C., area would have been less than ⅛ oz. over each square foot, not enough to produce an appreciable thickness. Much more is known to have been deposited over limited areas during some of the recent dust storms of the Plains area, but this deposition was due to exceptional and unnatural conditions. It has been estimated that Europe has received 5½ in. of dust during approximately 2,000 years (Free). This would be less than 0.07 mm. per annum, certainly not an imposing figure. There is little doubt that the rate of deposition of atmospheric dust is commonly so slow that ground organisms and rain are able to mingle thoroughly each deposit with the materials already on the ground and thus make impossible the preservation of stratification.

The thickness accumulated over a considerable period of time may attain significant magnitude. Deposits of loess in central United States attain a maximum thickness of about 30 meters, and in China thicknesses up to 90 meters are said to be present. Much greater thicknesses have been reported from China, but the reports seem to be in error.

In spite of the fact that deposits of dust from the atmosphere attain world-wide distribution, no wind-deposited dust of terrigenous origin has been proved to be present in any part of the Pre-Pleistocene geologic column. Deposits of volcanic dust are common.

Traction deposits may attain much greater thicknesses than deposits from suspension. Dunes 90 meters high are said to be present on the Bay of Biscay in France, and dunes of twice that height are said to occur in the Sahara Desert. The maximum height of the barkhan dune is given as 30 meters (Bagnold page 216), who states that the highest dunes appear to be in southern Iran where some seif dunes are said to have a height of 210 meters above their bases. The movement of dunes into low areas and hollows between ridges normal to lines of travel makes possible a great thickness. Tractional deposits have stratification, as deposition may be very rapid and ground organisms rarely live in the sands. Dune deposits cover extensive areas. There are 380,000 square miles of the Sahara covered with dunes, an area equal to a little more than one-eighth of the United States. The area of Arabia covered with dunes is placed at 320,000 square miles, and in addition there are extensive dune areas in western Asia about the Caspian Sea, in central Asia, in central and western Australia, in southern Africa, in western United States, Mexico, parts of South America, and there are also many hundreds of square miles in many parts of the world covered with sea-, lake-, and river-shore dunes. Dune sediments of the present surface rank among the major continental sediments.

Original colors of dune sediments are largely those of the constituent minerals, with the general range from white or gray to yellow. The sands of active dunes are usually white to yellow, but some dunes of tropical regions composed of lateritic materials are red to brown. Dusts from the atmosphere may settle on dune sands and impart to them the colors of the dust, so that light-colored sands may appear red. Secondary action may lead to other colors.

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TRANSPORTATION AND DEPOSITION BY ICE

Ice transportation has two aspects, depending upon whether the ice moves on land or is floated in water. Ice moving on land forms glaciers, and these make deposits directly; that floated on water gives up its burden of sediment to water on melting, and all fine material thus carried acquires the characteristics of aqueous deposition. Big pieces of rock embedded in ice sink on release and may introduce strange components into otherwise fine aqueous sediments.

Ice transportation on land differs from aqueous and atmospheric transportation in that there is no practical limit to capacity and competency so long as the load is on ice. If the load is near the base of the ice so that it has contact with the bottom, there is a limit to both competency and capacity which the ice solves by overriding the load. The loads of valley glaciers are often very large, and near the lower end of a valley glacier ice-borne sediments may be distributed from bottom to top. Loads of continental glaciers, as illustrated by the glaciers of Greenland, appear to be small. This arises very largely from the fact that valley glaciers have mountain slopes rising above them so that loads are acquired from above by sliding and falling on the ice in addition to that acquired by abrasion and quarrying on the bottom, whereas continental glaciers acquire little from slopes above them and obtain their loads largely by abrasion and quarrying from the surface over which they move. Loads are carried in and on ice and are pushed in front or dragged along the bottom. The former may be considered a form of suspension and the latter a form of traction. The limit to the tractional load is attained when it is easier to override it than to push or drag it along. In aqueous and atmospheric transportation the particles that travel by traction are larger than those transported at the same velocity in suspension. No such differences exist between ice suspension and traction, and, as a matter of fact, it is easier to transport large particles in suspension than by traction.

Most sediments carried by glaciers are deposited near places of melting of the ice. Some sediments are deposited beneath the ice at those places where the burden beneath is more easily overridden than pushed along. The *drumlins* are thought to have been formed in this way. This is direct deposition by ice, and the deposits are without stratification, sorting, or washing. The materials pushed in front of a glacier and deposited directly by ice at the ice front are also without these features. These form a part of the deposit generally termed the *frontal* or *end moraine*. As the end of a glacier holds different positions during recession, it results that an end moraine is formed at each place of halting. The moraine of farthest advance is generally known as *terminal;* others are *recessional.* As a glacier's front recedes, the sediments held by the ice may be directly deposited over the path of retreat to form a part of what is generally known as the *ground moraine.* This part is also without stratification, sorting, or washing.

Considerable parts of the sediments carried by glaciers are given to melt waters, which deposit their burdens as loads exceed capacity and competency. Channels beneath, in, and on ice sheets receive deposits which are wholly or in part held upon and within ice. These deposits are stratified, sorted, and washed when deposited. The channels and their deposits are subject to destruction as long as a glacier is advancing, but if parts of an ice sheet become stagnant, particularly if areas become isolated by severance from the main body of a receding sheet, the channels maintain permanent positions until melting has been completed. The deposits in the channels are gradually let down to the surface of the ground as the ice melts. They slump as this goes on, and slumping is completed when the ice has entirely melted. Stratification and sorting are destroyed, but evidence of washing persists. These are the *eskers*.

Any blocking of melt waters compels deposition. The first place where this is likely to take place is against the stoss side of a moraine, where melt-water streams flowing from a glacier are obstructed. Fanlike deposits known as *kames* form, of which one side is supported by ice. The ice melts and slumping is inevitable on that side. This destroys stratification and sorting; effects of washing persist. Stratification and sorting may be preserved on the other side. If a glacier advances up a valley, as was so commonly the case during the Pleistocene, a lake forms, of which the glacier is one shore. Melt waters flowing from the glacier have velocities checked on junction with the quiet waters of the lake, and deposition takes place in the form of a delta known as a *delta kame*. This is supported on one side by the glacier, and, as in a kame, slumping takes place on melting. A lake may also develop between a glacier and its moraine, particularly if there has been some recession of the front. Deposition in kames and delta kames is likely to be at high angles, and inclinations of 20 to 30 deg. are not uncommon. Deposits are made in lakes beyond delta kames. These are composed of fine sands, silts, and clays which may show seasonal or varved deposition.

Melt waters ultimately cross a moraine over low places and flow down the lee slopes. These strike lower surfaces and slopes, and, as they are usually provided with large loads, deposits of gravels, sands, silts, and clays are made in order with distance from the moraine, forming the assemblage of sediments known as *outwash*. These are stratified, sorted, and washed.

Thus, a glacial assemblage of sediments consists of a moraine of unstratified, unsorted, and unwashed sediments. This is flanked on the lee side by stratified, sorted, and washed gravels, sands, silts, and clays forming the outwash, and on the stoss side (and, in the cases of glaciers discharging into lakes, on top) by gravels and sands and rarely by silts and clays, forming the kames and delta kames. These stoss deposits are stratified, sorted, and washed adjacent to the moraine, but slumping has taken place on the side where they originally rested against the ice and has destroyed stratification and sorting. Locally and above the other deposits, there may be varved fine sands, silts, and clays, the deposits of icebound lakes. This assemblage of sediments is repeated at each recessional moraine and on a small scale over the entire path of recession of a glacier. Between moraines may be eskers and drumlins (Fig. 8).

The most striking feature of glacial deposits is the assemblage of sediments noted above. Detailed characteristics are as follows: As glacial sediments come from many terranes, they contain fragments of varied lithology. The strictly ice deposits are unstratified, unsorted, and unwashed and are mixtures of all dimensions and all degrees of angularity, justifying the name of boulder clay that is applied to them. Melt-water deposits exhibit rounding of particles from granule to boulder dimension. There is extreme interlensing of ice and melt-water deposits. The fragments are the results of physical forces, and there is no decomposition. The silts and clays are products of abrasion, grinding, and impact and are thus small particles of the original rocks and minerals. They are not the results of decomposition unless the original terranes eroded were composed of decomposed materials, which of course would be the case if glaciers invaded soil-covered lands. It has been suggested on theoretical grounds that glacially modified rock particles should approximate a flatiron shape (Von Engeln, 1930), but it seems far more probable that original shapes of rock particles have more importance in determining final shapes than anything else. Melt-water transportation tends to rounding, as under other conditions of aqueous transportation.

Glacially transported rock particles coming in contact with other fragments acquire percussion or chattermarks if direct blows are struck. These are no different from the same features developed by aqueous agencies. If rocks are rubbed over others, scratches known as *striae* are developed on both. This takes place to some extent within the ice but probably mostly where rocks in ice come in contact with the surface upon which it moves. Striated rock fragments are not particularly common, but striated surfaces over which ice has moved are not uncommon, and in many places the marks attain dimensions of grooves and furrows. These marks are developed by relatively low pressures that do not ordinarily exceed the crushing strengths of the rocks involved, and hence there is nothing resembling rock flow. The markings on the surface over which ice moves are generally in a single system; those made on blocks borne by a glacier may be in several systems and on more than one side, as a block may be rotated on a vertical axis to produce the several systems on one side and turned on some other axis to produce the striations on other sides. Striations are also made on fragments involved in earth and rock-slide movement, in mudflows, and also on the surface over which sliding and flow take place. The characters of these deposits, absence of stratification, washing, and sorting are also like those in the deposits made directly from glacial ice, but they may not have such wide variations in lithology, and the characteristic association of stratified and unstratified sediments of glacial deposits is altogether wanting. Striations may also be made by rocks carried in floating ice of any origin.

Floating ice has not a great deal of significance in rivers and small lakes. In large lakes the ice may transport large rock fragments from the shores and drop these in deep-water sediments, in which they form peculiar components (Von Engeln, 1918). The capacity of floating ice in the transportation of sediments depends upon the difference between the specific gravity of the ice and the water in which it floats, and therefore capacity is proportional to the quantity of ice. Competency, however, has no such ratio.

The ice of rivers, lakes, and the sea is derived from freezing of the bodies of water and from glaciers terminating in them. The former consists of shore or floe ice, and in polar latitudes (Kane) this ice is of great importance in transportation, as the sea remains more or less generally frozen and in places freezes to the bottom. Shore or floe ice picks up rocks from the bottom and shores and drops these on bottoms over which melting takes place. Sediments are washed upon frozen surfaces by streams that begin to flow before melting of the ice of shores. Sediments are also blown on ice by winds. The result is that considerable quantities of sedimentary materials are found in and on shore ice. The expansion of shore ice on warming may produce folds and overthrust faults which may serve to place bottom materials frozen in the ice above the bottom.

The great density of salt water gives to ice floating in the sea a greater capacity and competency, and as the ice masses are much larger than in other waters, there is opportunity for incorporation of large quantities and large masses of rock debris. Waves drive the ice masses upon each other and shove them into shallow water where they plow the bottom, disturb stratification, and pick up large volumes of bottom sediments and bottom organisms. If glacial ice masses are involved, the bottom may be modified for many feet from the surface. Some icebergs are more than 100 ft. high, which indicates that sea bottoms may be affected for a thousand or more feet from the surface. This is known to disturb sediments on bottoms of existing seas, and it must have done so in the past. Glacial ice entering bodies of water is ultimately floated, and the icebergs that break away from time to time carry large and small rock fragments derived from the land.

An interesting way for entrance of organic and inorganic materials into ice was discovered by the polar explorer Debenham. Water beneath floating glaciers and shore ice from time to time freezes and encloses the bottom organisms in the positions of life. At the same time the ice evaporates on the surface, so that in course of time it becomes sufficiently thin again to float and have water beneath. A new bottom population is then acquired. Again the bottom waters freeze, and thus the organisms previously incorporated are now above the bottom. Often-repeated occurrences of this nature may lift organisms entering the ice at the bottom to the surface, and an iceberg breaking away may be filled with organic remains from bottom to top. Debenham noted organic remains thus lifted 60 meters above sea level and 15 miles distant from the edge of the floating ice.

This same process accounts for the entrance into icebergs of many inorganic materials and also of sediments of which the origin is generally assigned to evaporation and arid conditions. Sea water on freezing expels its salts. This leads to concentration of the salts in the bottom waters; and, if circulation is impeded, these waters ultimately freeze and the salts are precipitated and ultimately frozen in the ice and brought to the surface in the manner outlined above for organisms. Debenham has noted a mineral (mirabilite) in this way in masses up to 0.33 meter thick.

Remains of organisms brought into floating ice in the various ways described seem at times to be in large quantity, as shown by Stefansson's statement regarding "bushels of small shells" in ice pressure ridges or Arctic seas.

Thus, ice floating in the sea and to a less extent in rivers and lakes may contain a large and varied assortment of organic and inorganic materials. This ice in course of time may float to warmer and deeper waters, there to melt and drop the sediments indiscriminately over all bottoms. This mingles large and small, angular and rounded fragments of rocks and organic materials derived from shallow waters of polar seas with sediments and organisms of deeper and oftener warmer waters. This is now taking place, and there is little doubt that it took place at times in the past. Unfortunately such features have rarely been recognized in marine deposits of the geologic column.

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TRANSPORTATION DUE TO CHANGES OF TEMPERATURE AND GRAVITY

Changes of temperature accomplish some movement that usually is downward. When materials are warmed, they expand in the direction of least resistance, which is generally downslope, and the succeeding contraction tends to create an equal degree of movement in the same direction. The movement is very slow, but almost daily there is some expansion and nightly some contraction. Like the movement of a glacier this movement can with difficulty be observed, but it is obvious in accomplishment.

Actual results due largely to changes of temperature may be seen in talus slopes, where, however, there is certainly some additional movement due to creep and slide. Davidson studied movement on talus slopes. He found the average rate of movement in summer with a daily range of temperature of 14.4°F. to be 0.04275 mm. per day, and in winter with a daily range of 8°F. the average rate was 0.03025 mm. per day. This is an average movement of a little more than 13 mm. in a year.

If the materials on a slope contain water and the region has freezing temperatures, each freezing of the contained waters lifts them in a direction perpendicular to the slope; each thawing results in a small downward movement perpendicular to the horizontal. The result is that the materials of the layer involved in the freezing and thawing have moved a slight distance down the slope. This, often repeated, ultimately brings the surface materials to the bottom of the slope.

Charles Darwin and Wyville Thomson described features on the Falkland Islands as "stone rivers." These had streamlike outlines and resembled glaciers in appearance. They were composed of very large blocks of quartzite that were supposed to have moved because of contraction and expansion. Bear Island in the Arctic Ocean has many "rock streams," and the island is extensively barren as plants cannot become adjusted to the moving soil. One "stream" described by Andersson had a width of 35 meters and a depth of 2.1 meters. The anterior margin of this "stream" through a width of 17 meters was composed of slabs of sandstone on edge parallel to the rounded margin of the "stream." This is probably a combination of slide, due to freezing and thawing, and flow. It was termed *solifluction* by Andersson. Capps has described features in Alaska that he termed *rock glaciers.* These were found in cirques and ranged in area to ¼ mile wide and 2½ miles long. Some had broad fan-shaped heads and terminated in narrow tongues, and others had pointed heads and fan-shaped tongues. Composing rock fragments were mostly small. The head parts usually had parallel longitudinal ridges, and the tongue parts concentric ridges parallel to the front. Rock fragments just beneath the surfaces were embedded in ice, and it was suggested that the formation and movement of these features were due to the freezing and thawing of ice. Ives's studies of "rock glaciers" in the Front Range of Colorado led to essential agreement with Capps's explanation of cause of movement. Tyrrell suggested that movement might be related to springs discharging into the debris.

Movement directly due to gravity produces rock or earth slides and falls, and creep and flow. In these movements the materials move en masse and not as individuals. The materials may be earth alone, rock alone, or a combination of earth and rock (Fig. 28). If the movement is earth and rock and the movement is slow, the large rock fragments tend to rotate backward about an axis parallel to the strike of the slope upon which the slide is taking place, so that, if more or less horizontal sedimentary rocks are present in the slide, many of those in the higher parts and perhaps in the whole of the slope dip into it. The Gros Ventre (Blackwelder, 1912) earth flow of the Lake Creek tributary of the Gros Ventre River in the mountains south of Yellowstone Park illustrates a slow movement. The materials involved consisted of soft clays and shales with some beds of sandstone and limestone. Movement could not be observed, but it seems to have been

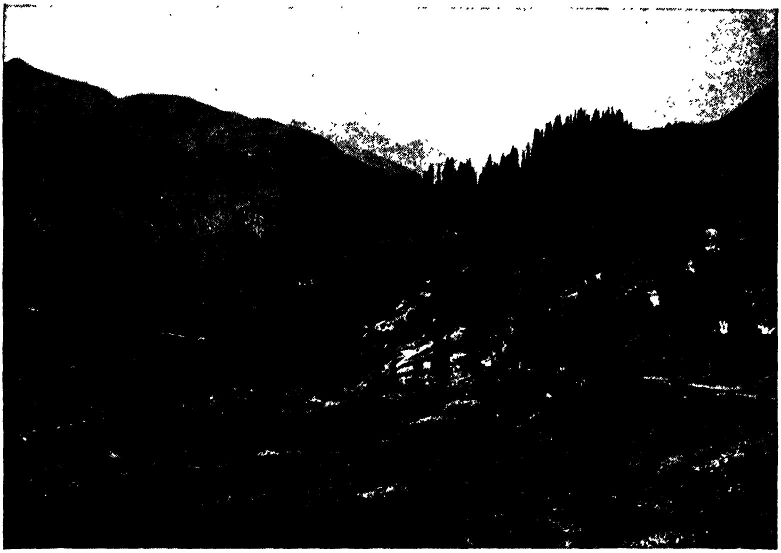

FIG. 28. Landslide deposit. Shows the hummocky surface resulting from this form of deposition. There is no sorting or stratification. Red Mountain Landslide District, Silverton Quadrangle, Colorado. (Photograph *by U.S. Geol. Surv.*)

continuous. Roads across the slide were displaced, and telephone lines could not be kept in repair. The movement was like that of a glacier, and crevassed domes developed about the lower end of the flow.

Rapid movement is termed landslide. This may be observed, and large masses may be involved. Earthquakes are frequently responsible for initiation of movement, and perhaps the best-known examples due to this cause are those that took place in the loess deposits following the great Chinese earthquake of 1920, in which it is said "the mountains walked." Movement ranged to ½ mile (Close and McCormick.)

Subaqueous sliding, or subsolifluction, has been observed on certain Swiss lakes with movement of large masses taking place on slopes of as little as 2 deg. 31 min. In the great Japanese earthquake of 1923 a block 10 km. long slipped to fill a depression to a thickness of 230 meters (Yamasaki) (see subaqueous slides in structures of sediments).

Rock and earth falls have the movement proceeding with great rapidity, as in the great slide at Turtle Mountain, British Columbia, in 1903, in which a mass slipped and covered a surface of 1.03 square miles to a maximum depth of 150 ft. (McConnel and Brock). The movement has been described as "a succession of great leaps or ricochets, probably accompanied by a certain amount of rolling and sliding." The rocks involved in such rapid movement are naturally badly shattered. A rotary movement is also developed, which, in contrast to the backward rotation taking place in slow movement, is forward, or becomes forward following first impact with any object, and, on landing, the shattered masses extend lobes of materials beyond the main body of the fall. These may be represented in the so-called rock streams (Fig. 29) in the San Juan Mountains of Colorado (Howe).

If the materials become saturated with water so as to resemble a thick viscous liquid, the movement becomes of the nature of a flow. The materials involved are mostly of small dimensions; hence the movement is termed mudflow (Pack). Earth mixed with water behaves as a viscous liquid, and the total distance moved may range from a few centimeters to several miles. Naturally a flow of great length creates the deepest impressions, but the small flows are the more numerous, and the quantity moved in small flows is probably greater in the aggregate than in large flows. The great Slumgullion (Howe) mudflow on the Lake Fork of the Gunnison River of Colorado started at an elevation of 11,500 ft. and flowed for 6¾ miles. The quantity moved was so great as to build a dam across Lake Fork and form Lake San Cristobal (Fig. 30). Blackwelder (1928) has studied many mudflows in the semiarid regions of the west and finds them to range in depth from a couple of centimeters to a meter or more with most in the range from 15 to 50 centimeters. Some are from 0.9 to 1.8 meters high on the outer margins.

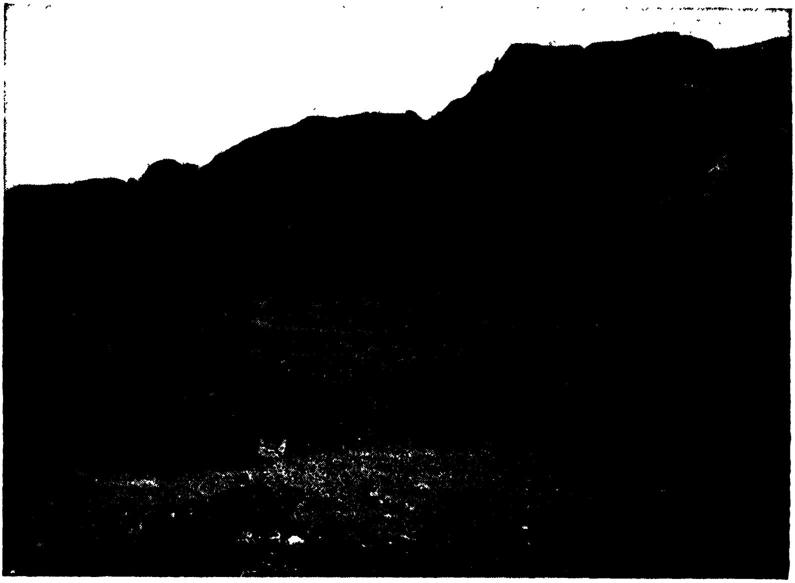

Fig. 29. Rock stream at head of Silver Basin, Silverton Quadrangle, Colorado. *(Photograph by U.S. Geol. Surv.)*

Fryxel and Horberg's studies of alpine mudflows in the Grand Tetons National Park, Wyoming, revealed that surface gradients of several large flows over the apical parts ranged from 7 to 15 deg. and to 3 deg. or less over lower margins. The flows terminated in abrupt slopes along the sides, commonly 2 to 6 ft. high and exceptionally 10 to 12 ft. high. Composing materials were unsorted and unstratified with the matrix light-gray mud or sand in which rock fragments and boulders to 4 ft. in diameter were embedded in large quantities at random. The largest flow studied was 700 ft. across and 500 ft. from the apex to the lower end. The flows ranged in thickness from 2 to 6 ft. in most and in a few from 10 to 12 ft. in marginal parts. Mud-and-rockflows have been rather carefully studied by the U.S. Forest Service over the past two decades in Utah, particularly over the Piedmont Region east of Great Salt Lake in an endeavor to discover causes. The Forest Service finds that the causes are overgrazing and burning over of the mountain slopes. A large mud-and-rock flow is shown in Fig. 31 (Bailey).

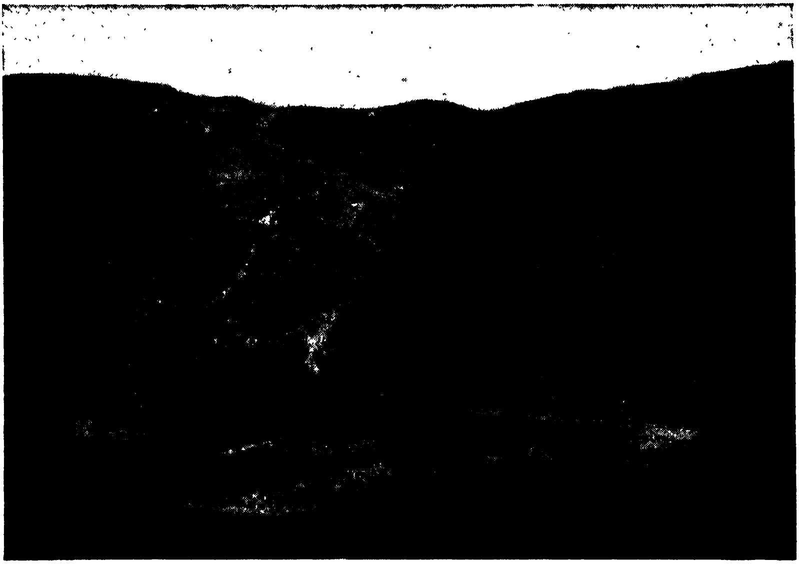

Fig. 30. The Slumgullion mudflow, San Cristobal Quadrangle, Colorado. The source of the flow is in the distance. Lake San Cristobal is on the right. (*Photograph by Whitman Cross, U.S. Geol. Surv., Prof. Paper* 67 (1909), 40–41, *Pl.* 20.)

Many mudflows, mostly small, follow thawing of frozen ground (Belcher). This is a movement of soil materials of great destructiveness on cultivated fields and otherwise bared surfaces. It is particularly prevalent in the spring over cultivated fields and in mountains of temperate latitudes. In mountains, streams of mud laden with rocks of various dimensions flow down mountain slopes, and in Patagonia the loose surface materials have been described as being in motion down slopes over surfaces previously polished by ice erosion (Coppinger).

Volcanic ash is likely to produce mudflows if water accompanies the falls or later wets them. Scrivenor describes one from Java that flowed for 38.25 km. and covered an area of 131.2 sq. km. with mud. The front of the flow ranged to 2.5 meters high.

Another form of mudflow is connected with the "bursting of bogs." Bogs by reason of capillary action may build their centers higher than the lands bounding them, and ultimately they "burst" and a mixture of water and black organic matter flows from the bog over lower marginal areas. A "bog burst" on the Falkland Islands back of the town of Stanley sent a stream of half-liquid peat through the streets of that town (Barkley). The stream was 1.2 to 1.5 meters deep and over 30 meters wide.

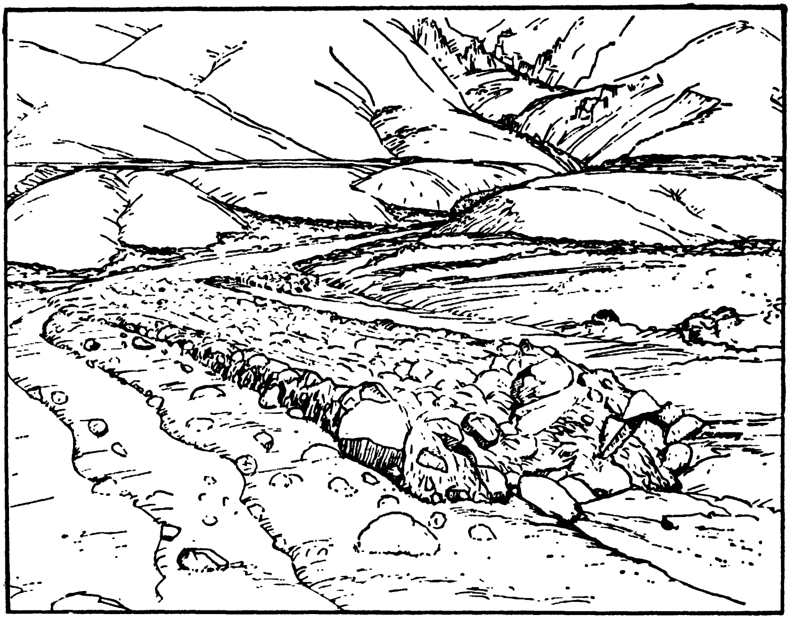

FIG. 31. Mud-and-rock flow in Davis County, Utah, 1939. Sketch made by the U.S. Forest Service in order to bring out more vividly the features of the flow. Instead of continuing down the axis of the alluvial fan when the apex was reached, the flow turned sharply to the left along the left margin of the fan. (*Courtesy of Reed W. Bailey, U.S. Forest Service.*)

Not a great deal of water seems necessary to cause mud to flow, and mudflows are more prevalent in semiarid regions, or regions that for any reason are without a cover of plant protection, than elsewhere. Vegetable growth serves to anchor the surface materials to deeper horizons and thus prevents flow.

The competency of mudflow to transport is extremely great, and blocks to 3 or more meters in diameter are readily transported. Flows in Utah are known to have transported blocks 12 to 15 meters long, and a mudflow at Willard, Utah, in 1924 is known to have carried one block with dimensions of 2.4 by 3.6 by 4.8 meters a distance of 7 miles. Flows of volcanic ash in Java carried large blocks of rock of which one was 3 meters high and more than 3 meters long.

According to Sharpe, the conditions that favor formation of mudflows are (1) . . . "abundant but intermittent water supply, (2) absence of any substantial cover of vegetation, (3) unconsolidated or deeply weathered material containing enough clay or silt to aid in lubrication of the mass."

Characteristic features of deposits made through sliding, flowing, etc., is the total absence of anything resembling sorting. Deposits consist of a jumble of large and small fragments ranging from clay size to blocks weighing hundreds of tons. Rock fragments are not rounded. Some may be striated, and striae may also develop on the surface over which movement takes place. Chattermarks may develop if movement is fairly rapid. Alluvial fans in semiarid regions contain many mudflows, and under these conditions they are associated with stratified deposits, although the stratification tends to be crude. This association may suggest glacial deposits, but not a great deal of discrimination is required to differentiate the deposits of the two forms of deposition, as the association of stoss- and lee-side deposits and central unstratified morainic deposits have no place in alluvial-fan and mudflow associations.

In spite of the fact that slide, mudflow, and like deposits are not uncommon in existing continental and, to some extent, aquatic deposits and must have been somewhat equally so in the past, there are few places in the geologic column where such deposits may be identified. An example of rock slide under aquatic conditions may be the Cow Head breccia in Ordovician strata on the west coast of Newfoundland. Here enormous blocks of rock containing assemblages of fossils ranging from Upper Cambrian to Lower Ordovician are enclosed in a matrix of Middle Ordovician age. The blocks seem to be of slide origin connected with an overriding thrust block pushed from the east. A similar origin has been postulated for the well-known Lower

Ordovician conglomerates about the city of Quebec, Canada, the blocks being postulated to have been derived from thrust blocks advancing northward (Bailey, Collet, and Field). A slide connected with fault movement has been postulated to explain large blocks of rock in Cretaceous strata on the west coast of Scotland (Dixon) and it has been suggested that the blocks of rock in the Caney shales of southeastern Oklahoma are slide blocks from a northwestwardly moving thrust block. Blocks in the Wildflysch of the Alps have been assigned to the same origin. No mudflows have been recognized in continental deposits antedating the present. This is probably because they have not been identified. It would seem that deposits of this origin might be present in such sediments as those constituting the Keweenawan of the Lake Superior Region, some "Red Beds" of western United States, the Old Red Sandstone of the British Isles, the Newark Series of eastern North America, and the Cretaceous and Tertiary continental sediments of the Rocky Mountains Region. Aquatic mudflow is thought to have been recognized in some contorted strata in the geologic column.

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ORGANIC TRANSPORTATION AND DEPOSITION

Organisms do considerable direct deposition of their protective and supporting structures and considerable indirect deposition in reworking already deposited sediments. Their main function seems to be placing sediments in such positions as to render them easy to acquire by water or wind, and in most instances the final agent of deposition of materials transported by organisms is one of these mediums.

Man (Fischer, Sherlock) is undoubtedly the most important of existing organisms concerned in the transportation and deposition of sediments, and one is reminded of the often-repeated comic picture or story of lost divers on the bottom of Lake Michigan who knew they were approaching Milwaukee by the increasing abundance of certain bottles strewn over the bottom. There is little doubt that in the far-distant future, when some present sea bottoms are bared, that geologists of those times, if the race still survives, will use various human implements for correlation of strata. Man's implements of culture and warfare and debris from his ships have been spread over all seas. His cultivation of land directly moves large quantities of material down slopes, and indirectly by removing vegetable protection (Glenn) he has greatly increased the quantity of sediments supplied to streams, and this has also greatly increased the quantity of water immediately entering streams. Floods are a consequence. There is little doubt that a geologic section through the deposits of the sea off the mouths of rivers draining cultivated lands would show the beginning of intensive human agriculture in the large thicknesses of mud thus produced. It has been stated by Sherlock that "in a densely populated country, man is about five times as effective as a denuding agent as all natural agents combined." If one considers the influence of man in erosion and hence in transportation of sediments in such mature areas as the Outer Blue Grass of the region of the Cincinnati Arch or the Driftless Area of Wisconsin, one cannot fail to be impressed that the removal of sediments due to human activity in these regions since the beginning of white occupation is such that it could not have been done by all other agents under the same conditions in a period at least fifty times as long. Hydraulic mining (Gilbert) during the past 50 to 75 years and the digging of canals, dredging of streams, and similar constructions have moved vast quantities of material. Hydraulic mining in California is estimated to have moved a billion and a half cubic yards.

Fossorial, or burrowing, animals move billions of tons of soil annually. Most of this is brought to the surface where parts are given to aqueous or atmospheric transportation. Among these animals are mice, gophers, moles, crayfish, muskrats, groundhogs, ants (Branner), and various worms. Earthworms under favorable conditions are estimated annually to bring to the surface a thickness of 5 mm. This amounts to over 700 cu. ft. per acre annually.

Plants directly move some material through growth of roots. This is not determinate, but probably that which is moved goes largely downslope. Aquatic animals also stir up the bottoms of the waters in which they live and thus indirectly become responsible for some transportation.

Floating vegetation is directly responsible for some transportation, and it is not an uncommon experience to see roots of trees so thoroughly grown around rocks that they cannot be released without cutting. If trees with rocks thus held fast are torn away by flood waters, the rocks are held fast until the roots rot or the tree becomes waterlogged. The rocks are dropped in the first place and are buried with the tree in the second. This may introduce peculiar components in certain sediments. It seems probable that the many boulders found in coal beds attained those positions in this manner. The tracing of these boulders to the distributive terranes should yield data with respect to ancient drainage, and each occurrence of such boulders should be recorded with all important data with respect to lithology (Ball, Spencer).

It is probable that from a quantitative point of view much important transportation is accomplished by marine aquatic plants. These grow most luxuriantly over rocky bottoms to which they are attached by holdfasts that firmly grasp objects on the bottom. These objects serve as anchors. Upper parts of such plants are commonly held up by buoyant structures that enable these parts to float on or near the surface. Many have numerous small such structures, but some have a single large balloonlike body with dimensions up to more than 10 cm. in diameter. Strong waves frequently tear these plants from moorings and float them to beaches or to shallow or deep water where they continue to float until the buoyant structures are punctured, or decay, when they sink and bring the "anchors" into deposits where ordinary aqueous transportation could never have brought them. The author has seen many hundreds of fragments held in the holdfasts of kelp on the beaches of the Gulf of St. Lawrence. Emery and Tschudy report a rock fragment weighing 13 lb. in the holdfast of kelp on a beach in southern California and a mass of gravel held in a holdfast which contained over 25,000 particles over 1 mm. in diameter. They also report the dredging of pieces of rock held in holdfasts from depths of over 700 ft. and observing kelp floating on the surface 250 miles from land. Many marine organisms make their homes on the seaweeds and the objects to which they are attached, and these are deposited where the plants decay or sink to the bottom.

Certain animals also function to buoy up stones. A ½-lb. stone, found on the coast of New Zealand, had evidently been floated by several specimens of an ascidian, *Boltema pachydermata*, which had attached themselves thereto. Growing upon the stalks of the ascidians were hydrozoans, bryozoans, and other organisms (Chilton).

Large-pored colonial corals may float almost as readily as pumice. Guppy has described one seen by him that had a circumference of 2.1 meters (7 ft.). It is not unlikely that such Paleozoic forms as *Favosites*, *Halysites*, *Lyellia*, *Heliolites*, and *Plasmopora* may at times have floated in the Paleozoic seas.

Many animals such as seals, many birds, some living reptiles, and the ancient herbivorous dinosaurs carry or carried rocks (gastroliths, stomach stones) in their gizzards. These introduce peculiar compounds into other sediments.

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DEPOSITIONAL INTERRUPTIONS

For many years there existed a more or less unconscious assumption that deposition is essentially continuous. Ultimately the great stratigraphic breaks known as unconformities were discovered. Later, stratigraphic breaks of a second order were recognized and termed *diastems*. Of later years it has been increasingly appreciated that deposition of continental and shallow-water marine sediments is probably more often discontinuous than otherwise and that the stratigraphic breaks in the known geologic column may represent a longer time than is represented by the sedimentary rocks. Study of continental deposits has demonstrated that numerous stratigraphic breaks are natural and common consequences of the conditions of deposition and cannot fail to be present.

It was at first generally assumed that each break in deposition was due to the surface of deposition having been brought above water level, and, as nearly all sediments were then referred to a marine origin, each break was interpreted as indicating a withdrawal and a return of the sea. As the stratigraphic breaks in continental sediments usually show much greater relief than those in marine sediments, it was frequently postulated that enormous thicknesses of sediments had been removed to create the unconformities. There was little or no appreciation that the relief is largely inherent in the manner of deposition. Within the past two or three decades it has been more and more appreciated that deposition is never continuous in continental deposits for a very long time and also rarely continuous in the marine deposits of shallow waters and that, over every shallow-water site of deposition, there have been many times when no sediments were deposited and also many times when the bottoms were eroded.

Deposition rests on two chief factors: a supply of sediments and a site of deposition fitted to receive deposits. Fitness of a surface to receive deposits depends upon the competency of the agents of distribution over that surface. If the supply at a given time and place exceeds the capacity and competency of the agents of distribution, deposition takes place; if the supply of sediments is equal to distribution capacity and competency, there is no deposition; and if distribution capacity exceeds the supply, there is removal of already deposited sediments. In bodies of water the distribution agents are the waves and currents, which vary in effectiveness because of winds, shore topography, and depth of bottom. Supply is determined by conditions at the sources of the sediments, which, as is well known, are extremely variable. The evidence and arguments for these interruptions of deposition, often accompanied by erosion of previous deposits, in the different sedimentary environments have already been presented.

Existing sea bottoms over the continental shelves seem for the most part to be below the level of the profile of equilibrium, so that deposition is raising these bottoms, but areas are by no means rare where the bottoms are stationary or are being lowered by erosion. In Chesapeake Bay about the mouth of the Choptank River (Hunter) the areas that have received no deposits during the 52-year period ending in 1910 are four times the areas upon which deposition took place, and even over the areas where deposition has raised the bottoms it is very probable that there may have been interruptions. This is a region where supply of sediments seems favorable for deposition. Agassiz states that the Blake Plateau beneath the Atlantic Ocean is receiving no deposits and contains few bottom organisms. The Indian Archipelago (Weber) has "hard bottom" in places at a depth of 1,500 meters, and "hard bottom" has been encountered in the Canary Islands at a depth of 2,000 meters. The bottom of the Strait of Gibralter is without sediments over much of its area. George's and Grand banks off the coasts of Maine and Newfoundland have Tertiary strata exposed on their surfaces (Verrill). The minimum depth is about 35 fathoms. The studies of Emery, Revelle and Shepard of the sea floor off southern California showed many places where rocks ranging in age from Triassic-Jurassic to Pliocene are not covered with sediments. Some of the places are 3,000 ft. below sea level, and most are banks and shelves that rise above the bottom. Revelle and Shepard, and Emery and Shepard have noted the existence of extensive areas of "hard bottom" over both shallow and deep ocean bottom off the California Coast. "Hard bottom" is known about the British Islands to the depth of 500 fathoms (Lebour), and Dangeard has found many places off the coast of France that are without sediments with rocks ranging in age from Permo-Triassic to Eocene forming the surface. It is very certain that there are many places on the present ocean bottom where sediments have not been and are not now being deposited and that many local unconformities due to nondeposition are in the making.

The fact that there exist bottoms over 2,000 meters deep that are not now receiving deposits suggests the extent to which such places may exist over shallow bottoms. As bottoms of this nature are now numerous, it is considered probable that this was the case to an equal degree in the past, particularly when broad shallow seas extended over vast areas of the existing continents and lands of the time were low.

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DIAGENESIS AND LITHIFICATION (ANDRÉE, SCHUCHERT, WALTHER, ZOBELL)

Diagenesis is defined as a term intended to include all modifications that sediments undergo between deposition and lithification under conditions of pressure and temperature that are normal to the surface or outer part of the crust and, in addition, those changes that take place after lithification under the same conditions of temperature and pressure, which are not katamorphic in character so that the effect is delithification. Changes of an anamorphic character that are due to temperatures and pressures which are not normal to the surface are excluded. Diagenesis is intimately connected with the relations of sediments to organisms, and many of the modifications antecedent to lithification are due to organic activity.

All aqueous sediments are saturated with water from the moment of deposition, and in fine sediments the water may constitute nine-tenths of the whole, and it is generally greater than half the weight. These waters invariably contain dissolved and colloidal substances, and there is no doubt that exchanges take place between the sediments and the permeating water and also between permeating and overlying waters. There is little or no doubt that there is much exchange of bases between materials in the waters and the sediments. In addition, the sediments are filled with organisms whose products add to the complexity of the mixture in the permeating liquids. The number of organisms is usually large even when casual observation indicates that few are present. MacGinitie states that in the mud of a slough of the Gulf of Monterey, California, he collected 974, with probably a couple of hundred uncounted, macroscopic organisms of 44 species in the muds taken from a circular excavation 45 in. in diameter and 18 in. deep. This amounts to approximately 170 macroscopic organisms per cubic foot, and, in spite of the number present, life did not seem to be abundant. Nothing was done with the microscopic organisms. These may have gone to millions per cubic centimeter. This mud contained the food and made the homes of this teeming population, and some of the mud at all times must have been passing through intestinal tracts.

As previously noted, it is thought that most marine-bottom sediments of slow deposition make many passages through intestinal tracts. These passages eliminate organic matter; change black, organically rich sediments to gray, organically poor ones; and the intestinal movements must produce some abrasion. Organic activities produce carbon dioxide, and some bacterial activities reduce sulphates to sulphides and ultimately form sulphur and sulphuric acid. The acid reacts with carbonates, and sulphates are formed upon which bacteria may again act. Some bacteria may precipitate carbonates. These may be returned to solution in waters rich in carbon dioxide and have the carbonates again precipitated by carbonate-precipitating bacteria. It seems likely that organic stabilizing colloids may be destroyed by bacteria, and colloids previously held in suspension may be precipitated. The results are that lithified sediments may bear little resemblance in color, composition, and texture to those first deposited.

Some results of diagenesis are expulsion of water, so that the sediments with depth contain less water than when first deposited and are more compact, and bleaching of red sediments to white through reduction of the ferric iron and formation of iron carbonate. Or the iron may be reduced and ferrous sulphide formed, with the result that the sediments become black. Lithification changes ferrous sulphide to the disulphides of iron, pyrite or marcasite, and the blackness is lost. Sediments black with organic matter may become light-colored through destruction of the organic matter, and this would take place where bottom waters have plenty of oxygen. The little-stable carbonate, aragonite, a common constituent of many shells and tests, is destroyed, and calcite may be produced. Under little-understood conditions some calcium carbonate is replaced by magnesium carbonate, and under other conditions some calcium carbonate is removed by leaching to enrich the sediments in magnesium carbonate. The products of such replacement or leaching in some way become dolomite. Other not yet fully understood reactions form glauconite, sedimentary feldspars, brookite, rutile, anatase, leucoxene, and certain zeolites like phillipsite, analcite and, possibly, natrolite (Merritt and Ham). There must be much base exchange between the sediments and the permeating waters. Volcanic ash alters to phillipsite, beidellite, and montmorillonite. Peat changes to lignite and finally to bituminous coal. The exact ways by which many of these changes take place are little understood.

The field of diagenesis is a big one and has hardly been touched. At one extreme in the formation of sedimentary rocks the raw sediments have been studied, mostly on the immediate bottom of water bodies and only recently for short distances beneath the bottom, and the lithified products have been studied at the other extreme. It has been assumed on the one hand that the lithified products represent the original sediments first deposited, which probably is rarely the case in fine sediments. It is probably the case for most sands and gravels except from the point of view of color. On the other hand it has been assumed that the freshly deposited sediments make rocks of the same composition, which again is probably rarely the case for fine sediments but may be true for sands and gravels.

The field of diagenesis invites research, but the problems are difficult and require training in, or cooperative efforts of, bacteriology, biology, and organic and inorganic chemistry. The problems of sampling also have not been fully solved.

Lithification is essentially a part of diagenesis, as chemical and physical reactions are set in motion which ultimately result in induration and cementation and recrystallization. When sediments are first deposited they are generally filled with water in which are various substances in solution and colloidal suspension. The quantity of water in fine-grained sediments is usually in excess of 50 per cent, and it is of the order of magnitude of 25 per cent or more in sands and gravels. Marls beneath a swamp in central Wisconsin at a depth of 11 ft. from the surface showed a water content of 69 per cent, and muds from the bottom of Crystal Lake in northern Wisconsin showed more than 70 per cent. Organic sediments from Lake Allequash at a depth of 16 meters beneath the bottom contained 90 per cent water. This water from the moment of deposition begins to be expelled, and compaction results, which is thought gradually to increase as the sediments accumulate. After the sediments have been buried to depths of 1,000 ft. or more, the water is thought to be largely expelled, and sediments originally with 50 per cent or more of water would decrease to half the original thickness.

Solution of sediments takes place where conditions are favorable. Chemical reactions between the substances in solution and the sediments take place from time to time, making new minerals that grow to fill empty spaces or increase the size of some of the old minerals.

Time is a factor, but not the deciding one, and sands, clays, and silts of the Cambrian are known that are as nearly unindurated and little cemented as they were on the days of deposition. It is not always clear, however, that the uncemented character of such sands is not due to decementation connected with circulating ground water. On the other hand, some Pleistocene outwash deposits are known that have become fairly well lithified.

Lithification is due to compaction, recrystallization, and cementation.

Compaction is largely due to the weight of overlying materials, which expels contained water and brings mineral particles closer together. This has its greatest effects on fine-grained sediments like silts and clays.

Recrystallization may begin in some sediments on the day of deposition. Permeating waters dissolve some minerals, and other mineral particles take from the waters the substances they require to continue growth, or new minerals develop as reactions to the new environment. Thus, a fine-grained lime mud may become crystallized to form an interlocking system of calcite or dolomite crystals, as illustrated by the Hoburgen "marbles" of Gotland or the Chicotte "marbles" of Anticosti. Silts and clays may become more or less recrystallized and indurated through the formation of crystals of illite. Quartz sands may become enlarged through deposition of silica upon them in optical continuity to form a quartzite of interlocking crystals of quartz. Gypsum may dehydrate and crystallize to anhydrite with a loss in volume of 38 per cent, or the reverse process may take place with an increase in volume of 61.3 per cent. The colloids and dissolved substances in the waters permeating sediments probably play a large part in crystallization, and much of each in course of time may be deposited in the interstices in the sediments. The colloids are commonly aluminum and iron hydroxides, silica, and hydrous aluminum silicates. These are in the form of gels after coagulation, and ultimately they lose water and harden like a glue or cement, thus bringing about cementation. It is thought that much of the disseminated chert in rocks may be due to this cause. Substances in solution may be anything, but commonly are calcium and magnesium carbonates. These substances are precipitated by chemical reactions, and carbonates also may be precipitated through loss of carbon dioxide. The dissolved substances were in the waters in the sediments when they were deposited, were derived from solution of some of the sediments, or were acquired from overlying or adjacent sediments or overlying waters. Sands of Bermuda dunes become cemented in the lower parts through calcium carbonate dissolved from higher parts of the dunes. Surface materials or semiarid regions frequently are cemented with substances derived by evaporation of waters rising from below through capillary action. Sandstones of humid regions are often case-hardened on the surface with substances in solution brought to the surface in dry weather through capillary action and deposited on evaporation of the water. Rocks when removed from quarries are said to be often much less indurated than after they have dried and the included waters have deposited their dissolved contents. Sand reefs on the coast of Brazil (Branner) are in places so strongly cemented with calcium carbonate as to resemble quartzites. This is postulated to have been done by deposition of the carbonates from stream waters as they seeped through the sands on their way to the sea.

Lithification is rarely the consequence of a single process. As compaction proceeds, there is solution, recrystallization, replacement, and cementation proceeding more or less simultaneously with the result that each product is different as dictated by the environmental conditions.

The most common cementing substances are calcium carbonate, silica, and iron oxide, considered important in the order named. Others are dolomite, siderite, pyrite, marcasite, various zeolites, and certain hydrous aluminum silicates.

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CHAPTER VII

CLASSIFICATION OF SEDIMENTS, SEDIMENTARY ROCKS, AND MINERALS OF SEDIMENTS

CLASSIFICATION OF SEDIMENTS AND SEDIMENTARY ROCKS

Any consideration of sediments and sedimentary rocks involves classification. As with most substances, this is difficult because of the fact of gradation from one form of sediments to another and the mingling of materials of different composition, provenance, and deposition by different agencies. Nevertheless, some classification is imperative, if for no other reason than that of convenience. Classification of any group of materials or organisms must rest on characters which authors consider basic, and these will vary with their training, personal experiences, and psychology. Different classifications have been proposed, and the bases suggested, or used, have included agents of deposition, chemical and physical composition, parts played by organisms, and environments of deposition. Several bases must be employed of which one must be given first rank and others subordinate rank. Grabau devised an elaborate system of classification of which each term endeavored to express chemical and physical composition, method of origin, and in some cases the environmental conditions under which the sediments were deposited. He proposed that rocks be placed in the two classes of *endogenetic* and *exogenetic,* the former including rocks formed as precipitates from solution, the latter, rocks formed from detrital materials. This would place the evaporites and granites in the same major group. This does not seem a reasonable association. Like most elaborate systems, this failed of adoption because of long usage of terms and systems already in use.

Pettijohn has recently advocated return to the Grabau basis of endogenetic and exogenetic genesis as the first basis of classification. This, as he states, places tuff and sandstone together in the exogenetic class and granite and rock salt in the endogenetic class. It is difficult to see where obsidian would go in this classification and also difficult to see how tuff could readily be separated from obsidian. The writer anticipates that granite will continue to be designated an igneous rock and rock salt a sedimentary rock with temperature of formation the basis for separation. Shrock has recently proposed a classification which has the merit of simplicity in that it follows the traditional classification and introduces only the three new terms of *silicastone* (flint, chert, etc.), *dolostone* for the dolomitic limestones, and *salinastone* for the evaporites. Krynine in his recent classification also follows traditional channels, using composition and texture as bases for classification and, like Shrock, he endeavors to retain old terms for which the writer thinks both are to be commended. Both recognize that there are all possible gradations from any form of sediments to most other kinds.

Through the years, sediments and sedimentary rocks have acquired many names, and it seems probable that most of these will continue to be used no matter what proposals or recommendations are made. Recognizing this as probable, the endeavor is made to build a classification with the names that are now in general use.

The sediments and sedimentary rocks are first classified on the basis of whether the agents of deposition were physical or chemical, thus giving two groups of sediments, one of physical and the other of chemical deposition. Sediments due to organic activities are placed with those of chemical deposition. Doubt must frequently exist, with respect to certain sediments, as to whether the agents of deposition were physical or chemical, as, for instance, shell matter on Bermuda that is heaped into dunes by winds, or carried by currents from shores and deposited in adjacent shallow waters. Sediments of physical deposition are classified on the basis of dimensions of particles with consideration given to composition through use of qualifying adjectives. Sediments of chemical deposition are classified on the basis of chemical composition. There are no sharp lines of separation, as most sediments of chemical deposition contain some materials of physical deposition, and the reverse is also true.

The sediments and sedimentary rocks are thus arranged as follows:

SEDIMENTS OF PHYSICAL DEPOSITION

a. The fine-grained clastics. Dimensions of $\frac{1}{16}$ mm. or less forming silt and clay.

b. The medium-grained clastics. Dimensions of $\frac{1}{16}$ to 2 mm. forming sand.
c. The coarse-grained clastics. Dimensions of 2 mm. or more forming granules, pebbles, cobbles, and boulders. These are generally known as gravel.

SEDIMENTS OF CHEMICAL DEPOSITION

a. Carbonate sediments.
b. Siliceous sediments.
c. Ferruginous sediments.
d. Carbonaceous sediments.
e. Evaporites.
f. Miscellaneous.

Assignment of sediments of chemical deposition to a class often presents difficulty in that some ferruginous sediments are carbonates, certain of the carbonates are evaporites, etc.

The various sediments pass laterally into each other, and any one may succeed any other vertically. Thus, a hole drilled in any particular area of a land may encounter every known variety of sediment, whereas in another area only a few varieties may be found. As a rule, however, not all varieties of sediments are encountered. The sediments found in a vertical column at any place constitute the geological column of the place. This column contains a record to which Marr has fittingly applied the name of *geogram.*

THE MINERALS OF SEDIMENTS

In any study of sediments two important considerations are the origin and character of the composing minerals and the use that may be made of the minerals for economic, stratigraphic, correlative, paleogeographic, or other purpose. The minerals may be placed in two groups, of which the individuals of one were derived from preexisting rocks. The minerals of this group may be of igneous, sedimentary, or metamorphic origin. These minerals are termed detrital or *allogenic* (*allothogenic*), and they may be termed *allogenes* or *detritals.* The minerals of the second group originate with, or in, the sediments of which they are a part. These are termed *authigenic* minerals or *authigenes.*

The allogenic minerals may be traced to their sources or distributive areas, and they are thus clues to provenances and environmental conditions by which they were released from the parent rocks. They aid to an understanding of ancient geography and climate. If the allogenic minerals are of value to human activities, they may assist in guiding to the mother deposits as shown in Fig. 32. Allogenic minerals may be used for purposes of correlation, particularly in subsurface work, where they may constitute the chief means to that end. They may also serve for the zoning of sedimentary sequences. In both of these uses there are serious limitations, and application should have full appreciation of these. A given mineral or a suite of minerals usually has application over only a single basin of deposition, and there should be no attempt of correlation from basin to basin.

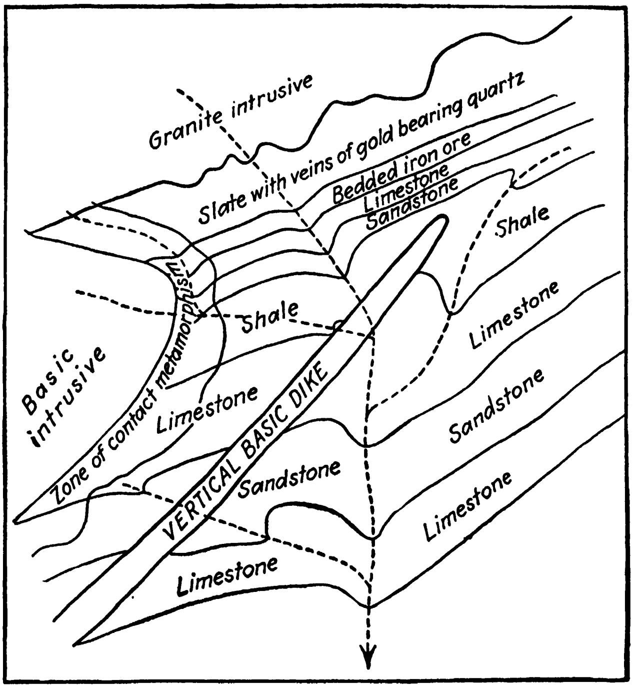

FIG. 32. Diagram showing distribution of minerals in fluvial deposits. The stream has four tributaries of which each may be expected to transport a suite of minerals different from the others. (*Modified after a diagram of Milner.*)

The allogenic minerals may greatly aid in an understanding of geologic history. The appearance of any particular mineral in a sedimentary deposit indicates that the parent terrane was in a position to supply this mineral. This may serve to determine the time when a batholith or other igneous unit was unroofed (Brammal, Groves). In California the appearance of glaucophane derived from the Jurassic Franciscan formation in the Lower Miocene shows that these rocks were exposed at that time, and as this mineral is not present in the underlying Kryenhagen formation (Tertiary) (Reed, Reed and Bailey), it indicates that the glaucophane-contributing rocks either were not then exposed or the derivative sediments were being transported to places where they have not yet been discovered. It has been shown in Wales that the basal beds of the Old Red Sandstone contain allogenic minerals derived from the much older Mono Complex, but that these minerals are wanting in higher strata of the Old Red Sandstone, thus strongly suggesting that the Mono Complex had become covered by the advance of deposition. Studies of the allogenic minerals in the Torridonian sandstones of Scotland have shown that they are younger than the Moine schist, which in turn is younger than the Lewisian gneiss (Mackie, 1928).

Climatic conditions may also be suggested by allogenic minerals. Under certain conditions, unstable and metastable minerals are decomposed to oxides, carbonates, and hydrates and thus disappear, and only stable minerals enter the sediments. The conditions that make this possible are known. On the other hand, there are conditions under which the unstable minerals are decomposed, but the metastable and stable escape. The conditions by which this is permitted are also understood. Thus, if unstable minerals appear in sediments, it may be postulated that climatic or other conditions obtained that made this possible. There are four possibilities: (1) the region of the parent rock was cold; (2) the region was arid; (3) there was such rainfall that products of rock destruction were not permitted to remain for a long time at places of production but were immediately transported to the sites of deposition; or (4) the sediments were broken from shores of seas, lakes, or rivers by the physical impact of water or were produced by grinding, abrasion, and impact of ice (Reed, 1928).

Regions with climatic and topographic conditions that permit mature decomposition are those of such moisture, temperature, and relief as to make it possible for decomposition to proceed for a long time, thus leading to the elimination of the unstable and, to a greater or less degree, the metastable minerals of the original rocks. Conditions leading to immature decomposition permit many unstable and most metastable minerals to persist, and an abundance in sediments of such minerals as feldspars, olivine, amphiboles, pyroxene, magnetite, and ilmenite indicates limitation of decomposition.

Allogenic minerals are derived from all kinds of rock. Each destruction of a sedimentary rock may eliminate some allogenic minerals originally derived from igneous or metamorphic rocks, and produce a relative increase in the proportion of stable minerals. After several cycles the allogenic minerals originally derived from igneous or metamorphic rocks consist only of those that are extremely resistant to decomposition or solution. Particles are also somewhat reduced from original dimensions. Sediments directly derived from igneous rocks are likely to contain stable, unstable, and metastable minerals, whereas allogenic minerals directly derived from sedimentary parents may contain no or few unstable minerals, few metastable minerals, and many stable minerals of original igneous or metamorphic parentage. The generalizations, however, must be qualified. The decomposition of an igneous rock may be so complete that all unstable and metastable minerals are destroyed, and the sedimentary parents of sediments may be arkoses or graywackes destroyed by physical processes without much decomposition.

Common allogenic minerals of igneous or metamorphic parentage are as follows:

Stable minerals:

- Anatase.
- Barite.
- Brookite.
- Cassiterite.
- Chalcedony.
- Chlorite.
- Chromite.
- Columbite.
- Corundum.
- Cyanite.
- Diamond.
- Garnet.
- Gold.
- Microcline.
- Monazite.
- Muscovite.
- Orthoclase.
- Quartz.
- Rutile.
- Sillimanite.
- Sphene.
- Spinel.
- Staurolite.
- Topaz.
- Tourmaline.
- Wolframite.
- Xenotime.
- Zircon.

Metastable to unstable:

Andalusite.
Augite.
Biotite.
Epidote.
Fluorite.
Hornblende.
Hypersthene.
Ilmenite.
Magnetite.
Plagioclase.
Pyrolusite.
Titanite.

Unstable:

Apatite.
Cordierite.
Glaucophane.
Olivine.
Pyrrhotite.

Some of these minerals, for example, diamond, are rare; others, as olivine, are common; and still others, as quartz, are abundant.

The chief authigenic minerals are:

Anhydrite.
Aragonite.
Barite.
Calcite.
Celestite.
Chalcedony.
Clay minerals.
Collophane.
Dahllite.
Dolomite.
Glauconite.
Gypsum.
Halite.
Hematite.
Limonite.
Muscovite.
Marcasite.
Pyrite.
Pyrolusite.
Quartz.
Siderite.
Wad.

Rarer authigenic minerals are:

Anatase.
Barite.
Brookite.
Celadonite.
Celestite.
Chamoisite.
Feldspars.
Native copper.
Galena.
Greenalite.
Leucoxene.
Malachite.
Pyrolusite.
Rutile.
Sphalerite.
Strontianite.
Sulphur.
Tourmaline.
Witherite.
Zeolites.
Zircon.

Any authigenic mineral may become an allogene. Several minerals are listed as allogenes and authigenes. Differentiation of origin may be difficult, but in most instances little difficulty may be anticipated.

The authigenic minerals may give some evidence of the conditions of deposition of the sediments of which they form a part and certainly of the environmental conditions that led to the development of these minerals. Unfortunately, knowledge of the developmental conditions of authigenic minerals leaves much to be desired. It is difficult to determine whether they formed at the time the associated sediments were deposited or long after.

Both groups of minerals may be divided into heavy and light fractions, the specific gravity of separation being that of 2.90, a heavy liquid of that specific gravity being used for separation. If the diameters of the particles are less than 0.05 mm., separation of the heavy minerals from the light is not readily made, as particles of that or smaller dimension tend to float indefinitely, even if the minerals are heavy. Separation of the heavy minerals from the light when the dimensions are this small may be made by centrifuge.

The use of minerals for correlative or other purposes requires recognition of all characters that the minerals possess. Thus, use is made of single minerals, suites of minerals, associations, sequences of minerals, percentages, surfaces, inclusions (Tyler), optical properties, and such physical properties as strain shadows.

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CHAPTER VIII

THE CLASTIC SEDIMENTS

INTRODUCTION

The clastic sediments include clays, silts, sands, and gravels. Classification is first based on dimensions and second on composition. Certain names call attention to origin. Each of the clastic sediments may originate in several environments.

Clastic sediments result from both physical and chemical methods of rock destruction, the former producing fragments and minerals of the parent rocks, and the latter particles and minerals of the parent rocks and secondary products resulting from decomposition. All kinds of rocks contribute. The particles have dimension, shape, and chemical and mineralogical composition contingent upon methods of detachment from source rocks; structures, textures, composition, and other characters of parent rocks; and history subsequent to detachment.

In some cases it is difficult to differentiate a rock of physical deposition from one of chemical or organic deposition, as illustrated by shell materials produced in one place and deposited by physical agencies in another, or a calcareous silt or sand produced by physical destruction of a limestone. The former is referred to chemical, the latter to physical, deposition. In general, sedimentary particles recognizable as detached from preexisting rocks are considered clastic sediments, whereas if deposited from solution or colloidal suspension they are generally designated chemical sediments; but argillaceous sediments carried as colloids are considered clastic sediments after deposition, whereas carbonate sediments carried in the same way form sediments referred to a chemical origin. This is not consistent, but there seems no way of escaping the situation.

Any clastic sediment is rarely entirely of physical deposition, as the sediments are commonly deposited in waters that contain much material in solution, and in these waters dwell organisms by which carbonates and other substances are formed. The result is that most clastic sediments contain a variable percentage of materials of colloidal and chemical deposition.

All particles of physical deposition have traveled greater or less distances. This travel has tended to eliminate the softer and chemically weak materials. Thus, long-traveled sands and larger particles tend to be composed of hard, tough, or chemically stable materials.

As materials are transported, they are subject to solution, abrasion, grinding, and additional breaking of the larger particles. The breaking tends to be controlled largely by such structural features as schistosity, stratification, and differences of texture leading to weakness. Transportation tends to place together particles of somewhat equal shape and dimension. Particles of spherical shapes tend to be rolled together, those of ellipsoidal shapes tend to become associated, disk-shaped particles tend to be sorted out from others, and particles of little rounding may lag behind in tractional transportation. It must be realized that sorting is rarely, if ever, perfectly done, and many deposits exist in which particles of all shapes are associated.

Table 17. Grades of Clastic Sediments

Limiting dimensions, mm.	Particles	Aggregate	Lithified product
> 256	Boulder	Boulder gravel	Boulder conglomerate
64 to 256	Cobble	Cobble gravel	Cobble conglomerate
4 to 64	Pebble	Pebble gravel	Pebble conglomerate
2 to 4	Granule	Granule gravel	Granule conglomerate
1 to 2	Very coarse sand grain	Very coarse sand	Very coarse sandstone
$\frac{1}{2}$ to 1	Coarse sand grain	Coarse sand	Coarse sandstone
$\frac{1}{4}$ to $\frac{1}{2}$	Medium sand grain	Medium sand	Medium sandstone
$\frac{1}{8}$ to $\frac{1}{4}$	Fine sand grain	Fine sand	Very fine sandstone
$\frac{1}{16}$ to $\frac{1}{8}$	Very fine sand	Very fine sand	Very fine sandstone
$\frac{1}{256}$ to $\frac{1}{16}$	Silt particle	Silt	Siltstone
$< \frac{1}{256}$	Clay particle	Clay	Claystone

The shapes of large particles are mainly determined by the nature of the parent rocks and the agents producing separation. It is considered probable that original shapes of large particles are largely responsible for ultimate shapes.

Table 17 gives the grades of the clastic sediments used in this work and the limiting dimensions (Wentworth, 1922).

Materials ranging in dimensions from 2 mm. to the largest possible dimension are classified as coarse-grained clastics and are termed gravel; those in the range from $\frac{1}{16}$ to 2 mm. are medium-grained clastics and form sands; and those in the range from $\frac{1}{16}$ to $\frac{1}{256}$ mm. (silts) and those smaller than $\frac{1}{256}$ mm. (clays) are the fine-grained clastics. In general, these terms are applied irrespective of composition, but the application cannot be made too rigorous, and if the particles in the range from clay to silt are composed of calcite or dolomite, the terms of silt and clay are not applied.

THE COARSE-GRAINED CLASTICS (HADDING, TWENHOFEL, WENTWORTH)

General Considerations

The coarse clastics are termed gravels, if unindurated, with a qualifying adjective designating dimension. After induration they are termed conglomerates, with the same qualifying adjectives.

The original fragments of gravels are produced by diastrophism, the various methods of rock breaking connected with weathering, and the considerable amount of rock breaking resulting from impact and grinding in transportation. Structures and textures of the original rocks and the agents responsible for the breaking determine the initial shapes of the fragments, and it seems probable that initial shape is the factor of greatest importance in determination of ultimate shape.

Most rocks by reason of schistosity, stratification, or the manner of breaking produce fragments with two large and one smaller dimension or two small and one larger dimension. Homogeneous rocks, such as granite, basalt, syenite, and gabbro, may on occasion produce fragments of approximately equal dimensions. After transportation, and the rounding more or less consequent thereto, the fragments of two larger and one smaller dimension form disk-shaped particles; particles of two smaller and one larger dimension produce ellipsoidal particles; and only fragments of approximately equal dimensions permit formation of more or less spherical particles. F. A. Fernald has suggested that rounded particles be termed *roundstones* to differentiate them from particles that have not been rounded.

Many particles of dimensions greater than pebbles may exhibit two or more generations of fracture, one that detached the particles from the parent rocks and others acquired after some rounding had developed. The latter may result from impact of two or more fragments, or crushing of smaller particles in the midst of larger. Impact has considerable effectiveness on brittle rocks, but is relatively ineffective on such tough rocks as basalt. Rocks with structural planes of weakness may also break easily on impact. Frost action operating along lines that absorb considerable water may also cause breaking.

Under the conditions of arid and semiarid regions, but also in regions subject to seasonal dryness, particles may acquire a coating known as desert varnish (Laudermilk). Frequently rounded particles are found in iron oxide deposits that have a coating resembling desert varnish. These are common in the upper Ordovician iron ores of eastern Wisconsin and consist of clay with iron oxide coating.

Surface parts of gravels are often more firm than the interiors. This is supposed to be due to the removal of cement from the interiors and is termed *case-hardening.* Case-hardening seems to be most effective in sandstones and limestones.

Gravels may be composed of any kind of rock. Most are composed of tough and hard rocks, and this is most likely to be the case if they have undergone long transportation. Gravels composed of soft or brittle rocks are not likely to travel far from distributive areas unless the transportation is accomplished by icebergs or glaciers. Gravel composed of clay (Haas) is not uncommon, and the particles tend to grow larger by accretion as they are transported, and they may become studded with pebbles, granules, and/or sands. Even so fragile a substance as peat forms gravels with particles in the range from granules to boulders. Bones are not uncommon as gravels in the range from pebbles to cobbles. Many products of industry are found, as rounded particles of glass, brick, or concrete, and on some coasts particles resembling gravels in the range from pebbles to cobbles are composed of wood.

Transportation involves wear and hence decrease in dimension, so that after long transportation the larger coarse clastics tend to change to small ones and ultimately only small particles remain. Thus, large particles, such as boulders, indicate nearness to distributive areas unless the transportation was effected by ice. Cobbles are less committal, as the original fragments may have been of that dimension in the beginning, but if the cobbles are composed of very hard or tough rocks, long aqueous transportation is a reasonable postulate. The same assumptions are possible for the smaller grades of gravel.

Surfaces of gravels transported by water tend to be smooth, but more or less greatly marred by chattermarks and the marks of abrasion. Gravels transported by ice may have the surfaces unchanged, but if gravels are in the base of a glacier, they may frequently contact the rocks beneath to develop facets, grooves, and striae. Gravels left behind through removal of finer materials are termed *lag gravels.* These may have etched and polished surfaces with strong minerals in relief.

The rate of wear of gravels tends to be proportional to the weight of the particles in water and the distance traveled. Wentworth defined as *reduction index* the loss in weight in parts per thousand per kilometer of travel (1931).

Sorting of coarse clastics has wide variation and is best developed in the finer grades and is ordinarily very poor in the coarse grades, since the interstices between the large particles are so great that particles of a wide range of dimension may be lodged therein and be protected from removal. Among the finer grades sorting may attain an apparent high degree of perfection as seen by the eye.

The stratification of gravels is rarely excellent, and in very coarse gravels it may be almost entirely absent if one considers a small section, but if the entire gravel unit is considered the bedding may become apparent. A gravel unit is usually more sharply defined at the base than at the top where there tends to be gradation into the overlying stratum. Such gradation may also be present at the base, and there may be a sharp change at the top. Gravel beds may be cross-laminated but such is rarely apparent in deposits composed of cobbles and particles of larger dimensions. Ellipsoidal gravels deposited by water on sands tend to lie with the long axes perpendicular to the direction of flow of the currents to which they owe their deposition. If deposited with other gravels, positions of long axes may be various.

Disk-shaped gravels of current deposition tend to be imbricated with the upper surfaces of the particles inclined up-current. Gravel of similar shapes deposited on beaches tend to imbricate seaward (Fig. 24) with the upper surfaces inclined in that direction (Johnston, Pettijohn).

Corals, algal structures, and concretions have shapes and dimensions like those of gravels and may be mistaken for them. This is not so likely for corals, but is not unlikely for algal structures, as these do not show organic structures unless they are broken. The organic structure may then become apparent. Concretions should also be broken to show the concentric or radiate structure commonly present.

There is great variation in the cementation of gravels, and the cements are of various kinds. Usually smaller particles, such as granules, sands, silts, and clays, fill interstices between larger ones, and calcite, silica, clay or some form of iron oxide unites the whole. Glauconite in some cases assists in filling the interstices. The cementation in some cases is so excellent that fracture may take place across particles, and in others so poor that the gravels may be readily removed with pick and shovel or steam shovel. Indurated gravels form conglomerate.

It has been taught as almost axiomatic that gravels are invariably associated with unconformities, and *basal conglomerate* is an ingrained concept in the literature. The validity of this concept is extremely questionable, as in many cases there are no gravels above an unconformity and in other cases the gravels are below and not above. Moreover, there are many gravels that have no relation to unconformities. In many cases the gravels are marginal to the units of stratification and not basal.

Gravels are deposited in streams, on lake and sea beaches, in shallow waters of lakes and the sea, by glaciers and their melt waters, by icebergs, and are left behind by winds and waters as they remove the finer materials within their competencies (Mansfield).

The sorting of river gravels (Gregory) ranges from poor to excellent. They may be poorly to well rounded, the former the more common, and rounded individuals range from disklike to ellipsoidal and spheroidal. If river gravels lie stationary for a considerable interval of time, they may be abraded by passing particles upon the top and up-current sides, and there may be more or less etching. Over the piedmont (Lawson) parts of the fluvial environment, sorting, rounding, and stratification tend to be poor, and composition is apt to be varied. Beds are rarely continuous. In lower reaches of streams there is greater homogeneity of composition, as some of the weaker rocks have been eliminated, sorting is better, and there may be greater continuity of bedding. The thickness of fluvial gravels with a stationary crust is limited, except over piedmont areas, where it may be great. If a stream flows through a subsiding area, a great thickness is possible over such portions of a stream's depositing area.

Beach gravels (Gregory, Marshall) range from poorly to well sorted, and, as a general rule, sorting tends to be fair to good. Stratification is extremely irregular and more often wanting than present. Particles may be angular to well rounded. Shapes range from ellipsoidal, disk-shaped, and spheroidal to highly angular. Shapes consistent with specific gravity tend to become segregated, so that some beaches have more or less spherical gravels and others disk-shaped. Landon's studies indicated to him that angular fragments of beaches change from angular to spherical and then tend to become flattened. It is the writer's observation, based on travels on foot over fully 1,000 miles of beach on the coasts of Newfoundland, Anticosti, Labrador, Quebec, and elsewhere that the disk-shaped gravels of beaches are due mostly to structural features of the parent rocks that led to the original fragments having two large dimensions and one smaller one. The tendency for gravels of these shapes to be segregated may be referred largely to a source in the immediate vicinity contributing fragments of the necessary shapes or to selective sorting. On many beaches disk-shaped particles are few or wanting, and the particles range from ellipsoidal to spheroidal.

Beach gravels may have extensive distribution in a single direction, but the distribution in a direction normal to the shore is limited. The thickness is small and rarely exceeds 6 meters. It usually is much less.

Offshore gravels of lakes and the sea tend to have ellipsoidal to spheroidal shapes because of the fact that it is relatively easy for these shapes to be rolled outward from shores. They tend to be interlensed or dovetailed with sands and finer sediments. Beds may extend over extensive areas with fairly uniform thickness. Barrell expressed the opinion that marine gravels are not carried to depths greater than 20 to 30 fathoms or to distances from the shores of 10 and in a few cases 15 miles. Observations made on the Anticosti section show that gravels are present in the deposits of that section that could not have been deposited closer to shores than 30 miles. The Ordovician and Silurian of Newfoundland contain conglomerates of marine deposition that are hundreds of meters in thickness, and fragments from the cliffs on the south coast are now being deposited in deep water to depths of 300 meters. It is conceivable that a sea rising over a youthful land would lead to deposition of gravels in deep water with a near source generally present as the water rises, and concomitantly with the rise of the sea additional gravels would be deposited above those laid down during the time of the previous level, so that a large thickness might be possible. The units would be in the form of wedges extending into finer deposits made in the deeper water offshore.

Glacial gravels tend to be subrounded to angular except in melt-water deposits of which the gravels may be as well rounded as the gravels of any aqueous deposit. Ice-deposited gravels have abrasion facets produced by rubbing of fragment over fragment or over bedrock. Some fragments may be striated or grooved, but such marks are not so common as is generally supposed, and little extraglacial transportation serves to obliterate striae. Many ice deposits have rock fragments with the shapes the same as on separation from the parent rock. Glacial gravels may attain thicknesses as great as 300 meters, but, ordinarily, common maximum thicknesses range from 30 to 60 meters. Those deposited directly by ice exhibit no sorting or stratification and contain much clay, silt, and sand. The unsorted parts dovetail with melt-water deposits that are sorted to a greater or less extent (Von Engeln; Wentworth, 1936).

The lag gravels left behind by wind exhibit varied degrees of rounding, and, if in the paths of drifting sands or silts, the corners are more or less rounded. Many have facets. These may be polished original fracture surfaces, or they may have been developed by sand abrasion. They are usually not so flat as the facets on glacial gravels. Most are thought to be polished and etched original fracture surfaces. If there are considerable differences in hardness of the composing minerals, the hard individuals stand in relief. Hard, homogeneous rocks like chert or flint may have the surfaces polished to a mirrorlike finish. The particles so modified by an aeolian action are known as *ventifacts*, *sand-blasted* gravels, or *einkanter*, *zweikanter*, etc., of the Germans (Bryan).

The thickness of lag gravels can hardly be great unless the aeolian agencies are acting over a gravel deposit in process of deposition by other agents of deposition, with deposition proceeding at a rate slow enough to allow the aeolian agencies to produce the characteristics of ventifacts upon the gravels as they are deposited. Under such conditions a great thickness may be possible. Aeolian lag gravels tend to form in greatest abundance over arid and semiarid regions, but they may form wherever there is sufficient absence of vegetation to permit effective deflation, as on high mountains, in high latitudes, and on sea and lake shores.

Residual gravels have shapes ranging from more or less spheroidal to highly irregular. Dimensions range from very small to large. The surfaces may contain corrosion pits of greater or less depth. Residual particles tend to exhibit spheroidal weathering except in the cases of flint and chert, where this may not be marked. Thicknesses may attain 6 meters or more.

Boulders and Cobbles

Boulders range in dimension to several meters in diameter. Large boulders tend to be poorly or not at all rounded. Rounding is better developed as the minimum limit is approached, and on some beaches the rounding may have become so excellent as to have obliterated original shapes. The largest boulders are transported in mountain streams, by waves and currents on stormy coasts, by glaciers and mudflows on land, and by icebergs in seas and lakes.

Well-rounded boulders are present most commonly on beaches, but they may also be present in the lower parts of mountain streams and in glacial outwash. Spheroidal and ellipsoidal boulders are generally composed of homogeneous rocks. Large boulders are rarely well rounded.

Rounding is mainly accomplished by abrasion and to some extent by impact, both connected with the rolling of particles. It may be done in a short time (Bartrum). Boulders that become fixed in position with finer sediments moving over them may become flattened or grooved by abrasion, and if the composing materials are of different degrees of abrasive resistance, the surfaces may become more or less etched with the hard minerals in relief. Boulders thus fixed in position constitute a natural boulder pavement.

Boulders may be composed of any kind of rock, and it seems probable that more kinds of rocks may be found among boulders than among any other coarse-grained clastics. Boulders are not necessarily composed of hard and tough rocks. Boulders of clay are not uncommon, and boulders of peat are known.

Cobbles are generally better rounded than boulders, generally have wider geographical distribution in any environment, and generally have about the same range of composition as boulders. They may have averaged longer travel than boulders and hence average a greater percentage in composition of resistant, hard, and tough rocks. English geologists frequently describe the coarse gravel of beaches as *shingle.*

Pebbles and Granules

Pebbles and granules may have their respective dimensions because of original size on detachment from parent rocks, but they may also have become pebbles or granules because of long travel and much abrasion, impact, and grinding. They tend to be better rounded and more spherical than cobbles and boulders, and some attain a high degree of sphericity. Pebbles and granules are more often composed of resistant rock than cobbles and boulders. Both have a wider geographic distribution than the more coarse clastics, and they are present in fluvial, lake, marine, desert, glacial, and, occasionally, other environments.

Edgewise Conglomerates

This type of conglomerate has the rock particles in some position transverse to bedding, and the range in position in a hand specimen may be from normal to parallel to bedding. This form of conglomerate—many are breccias—has extensive distribution in the Upper Cambrian of the upper Mississippi Valley, the early Paleozoic strata of the Appalachian Region, and in the Deadwood formation (Ordovician) of the Black Hills.

No very satisfactory explanation of the origin of edgewise conglomerates has been given. They have been referred to rapid deposition, to slumping after deposition, and to breaking up of previously deposited strata before complete induration followed by later rapid heaping together of the fragments. Beds of this type of conglomerate are known in the Upper Cambrian of Wisconsin that have horizontal distribution in horizontal beds over many miles. Such distribution renders impossible any explanation due to slumping. It is probable that this type of conglomerate may develop in each of the ways postulated.

Intraformational and Interformational Conglomerates

Walcott defined *intraformational conglomerates* as those formed from the breaking up of a partially indurated stratum and the aggregation of the particles without much transportation. Unfortunately the illustrations given by him as representative do not seem to have been formed in the manner postulated. The term, however, is a good one and should be retained, as conglomerates originating in the way postulated are thought to exist in the geologic column (Field). The term of intraformational breccia is often more applicable. Conglomerates or breccias of this origin may develop where mud-cracked polygons become broken up by strong and rapid water action.

The term *interformational* is applied to those gravels and their indurated equivalents that often are present within a formation of which the constituents have a source external to the formation. These conglomerates may be found in association with any kind of common sedimentary rock and merely record times when the transporting agents had more than ordinary competency.

Till and Tillite

Till is an unstratified and unsorted aggregate resulting from ice deposition. Tillite is the lithified equivalent. The physical composition ranges from particles of extremely small dimension to blocks of many tons' weight. The composing materials are mainly of physical production except in the glacial deposits of farthest advance, in which there may be much of decompositional origin. Fragments tend to be little rounded, and many have abrasion facets. Some may contain striae, but such marks are not common, and, moreover, striae may be produced in other ways. The rock fragments as a rule show little decompositional effects, but deeply corroded fragments are not uncommon. In any locality most fragments are of local origin.

Sedimentary Breccia

Sedimentary breccias (Field) are composed of angular fragments of gravel dimension derived by sedimentary processes from preexisting rocks in distinction to pyroclastics and autoclastics (dynamic and tectonic breccias), the former composed of volcanic fragments and the latter of particles produced by fracturing and faulting. Several important varieties of sedimentary breccias are talus, bajada, mudflow, desiccation, and residual breccias (Norton).

Talus breccia or scree material consists of angular, small to large fragments of rocks that are derived from, and accumulate at, the foot of cliffs. The cliffs may be on land or border a lake or the sea. The material consists of angular, unsorted, and unstratified rock fragments of a wide range of dimension. The finest material tends to be at the top of the talus slope, but decompositional products may be more abundant near the base.

Rock glaciers, mudflows, and landslides may form breccias that may not be greatly different from talus breccias except that, incidental to a slightly greater distance of transportation, there may be a little more variety in the kinds of fragments and mud would be present in flows. Some fragments may be striated or faceted, the latter not likely.

Bajada accumulations are somewhat similar to mudflows except that more water is involved in transportation and deposition, and hence there may be a little sorting, stratification, and rounding of the particles. Formation takes place in arid regions from adjacent mountains where coarse clastics, mud, and water flow in streamlike fashion down the slopes (Woodford). Bajada breccias grade into typical mudflow breccias.

Desiccation breccias may form wherever there is much mud cracking. The mud-cracked polygons may break into fragments which later may be deposited with other sediments.

Glide breccias form in subsolifluction on sea and lake bottoms through sliding of sediments deposited under somewhat unstable conditions at higher levels. Some of the so-called edgewise conglomerates may be breccias of this origin.

Residual breccias result from concentration of rock fragments from washing or blowing away of associated fine residual or deposited materials. These are some of the so-called lag gravels of deserts and the coarse, residual particles that mantle many of the deeply eroded lands of the United States and elsewhere. The *gibbers* of the Australian deserts are of this origin. Much residual breccia is composed of flint or chert and in some localities is so abundant as to render areas nearly valueless for cultivation.

The Dinantian of Belgium (Mississippian) contains a breccia in marine deposits about which there has been considerable controversy among European geologists. The thickness is several tens of feet, and there are two facies, one of red color, the other gray. The gray breccia is the more extensive. Some consider the two of the same origin, others consider them different. These Dinantain breccias have been studied by Dordolet and by Kaisin. The latter gives references to other studies. The author has no opinion respecting origin.

THE MEDIUM-GRAINED CLASTICS

Sands and Sandstones (Allen, Dapples, Hadding)

The medium-grained clastics are sands and their indurated equivalents, sandstones. Sands are composed of particles of dimensions between $\frac{1}{16}$ and 2 mm. which were derived from preexisting rocks. The last qualification leads to exclusion of shell fragments, oölites, and volcanic ash of sand dimensions.

Sands and sandstones ordinarily have the majority of the particles in a closer range than $\frac{1}{16}$ to 2 mm., but there are many examples in which the entire range is present, as well as particles beyond the limits of the range. It seems probable that there are far more sands and sandstones that contain important contents of silt and clay than those containing important contents of granules and pebbles. This may be more apparent than real, however, since with rise in content of granules and pebbles the tendency would be to term the aggregate a gravel or a conglomerate rather than a sand or sandstone.

The mineralogical composition of sands has wide range, and there are probably no varieties of the common rocks and minerals that are not present in sands in significant quantities. The most common constituent is quartz, and many and probably most sands are dominantly composed of this mineral. Sands that have undergone long transportation tend to be largely reduced to this mineral through elimination of the less resistant constituents. Other minerals commonly found in sands are zircon, garnet, tourmaline, and less commonly rutile, titanite, cassiterite, andalusite, staurolite, monazite, spinel, and topaz. Some sands are composed mostly of ilmenite, magnetite, chromite, or garnet, or some combinations of these minerals. Other sands are composed mostly of feldspars; others have a large content of muscovite or biotite; and sands composed of limestone particles are extremely common on the beaches of the islands of Anticosti, Gotland, and other places where limestones form coastal cliffs. Gypsum or anhydrite forms sands over a considerable area in New Mexico (Herrick). Epidote, amphibole, and pyroxene are not uncommon in some sands. Coral "sands" and "sands" produced by the breaking of shells compose the beaches of many tropical regions, but as these are not derived from preexisting rocks, they are not placed with the clastic sediments. Black sands dominantly composed of particles derived from Keweenawan lavas form many beaches of the Keweenaw Peninsula and other places on the shores of Lake Superior. Black sands on the coast of Oregon are composed of quartz, olivine, epidote, garnet, magnetité, ilmenite, chromite, and zircon. These sands are local concentrates of complex history. The largest grade size is mostly quartz, the next to the smallest is mostly magnetic minerals and zircon. Below $\frac{1}{16}$ mm. the mineral materials are clay and quartz (Figs. 33 to 36). Beaches marginal to the places of local concentration of the black sands are composed largely of quartz, and the grains are less well rounded than quartz grains associated with the black sands (Fig. 37).

Many sand grains, particularly those composed of quartz, contain inclusions. These may be an excellent guide to parent rocks, and the original parents may be determined even after the sands have passed through several cycles of erosion and deposition (Tyler).

The shapes of sand grains range from highly rounded and spherical to very angular (Anderson, Cox, Sherzer, Twenhofel, Ziegler). Aqueous sands show little rounding on particles with diameters below about 1/10 mm. This is readily shown by separation of any sand into fractions based on dimension. The fractions of fine particles are almost entirely angular. Wind-transported sands may be rounded to lower dimensions than can be done in aqueous transportation. Galloway has stated that a noncalcareous sand with more than 50 per cent of the particles well rounded is more likely of wind than water production. He placed the lower effective limit of abrasion by water at 1/20 mm. and for wind at about 1/30 mm. Although this may be the theoretical lower limit for water-borne particles as shown by controlled

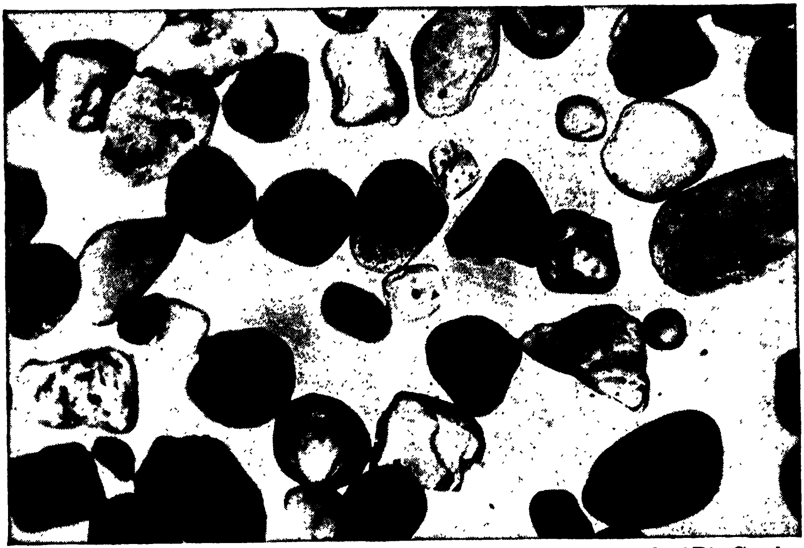

FIG. 33. The 1/4–1/2 mm. fraction of sand collected 1,000 feet north of Big Creek, Agate Beach, coast of Oregon (× 55). The light-colored grains are quartz of which many contain inclusions, some of which are magnetite. The well-rounded grains are mostly olivine and epidote; a few are pieces of basic igneous rock or black shale. There is an occasional grain of magnetite. This sample is one of rich concentration of magnetic and heavy minerals and the rounding of the grains has reached a higher degree than in sands of the same beach where there has been no concentration of heavy minerals and the particles are predominantly quartz. Photographed in liquid of index 1.63. (See Figs. 34 and 35.)

laboratory experiments, the limit seems rarely attained in natural waters, and few quartz particles below about 1/10 mm. are much rounded in aqueous transportation. Abrasion in the intestinal tracts of organisms and solution may round particles to very small dimensions, and the lower effective limit from these causes must approach colloidal dimension (Kindle).

It is the author's considered opinion that rounding of sand grains, particularly of the dimensions of 1/2 mm. or less, involves extremely long tractional transportation. Observations of sands of such dimensions which were brought to the sea by the Columbia River of the Northwest showed essentially no rounding after travel which at the very least must have been several hundred miles, and sands on the beaches of Washington and Oregon and in the coastal dunes showed grains to have undergone negligible rounding so far as the grains of quartz were concerned. These sands must have been moved around on those coasts for centuries.

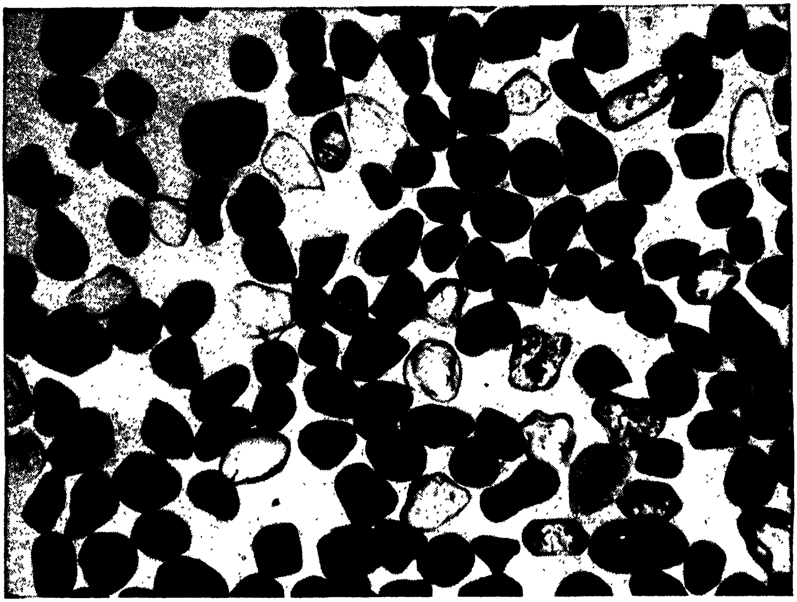

FIG. 34. The 1/16 to 1/8 mm. fraction of the same sand as Fig. 33 (× 64). The fairly well-rounded grains are magnetite, ilmenite and chromite. These grains have shiny, polished surfaces. The angular light-colored grains are quartz or garnet, the former containing inclusions. The ellipsoidally rounded grains of high relief are zircon. Photographed in liquid of index 1.63. This fraction contains little quartz and is predominantly composed of black minerals.

To produce well-rounded grains of quartz of 1 mm. or less in diameter required not a few hundred miles, but many thousands of miles of travel.

Visual impressions of sphericity and rounding do not have high value. It seems probable that sand particles are less spherical than pebbles and granules. It also seems that sands of aeolian transportation are more spherical than those of aqueous transportation, but examples of excellent sphericity in aqueous deposits are not uncommon, and sands of aeolian transportation are very frequently angular. Glacial sands are likely to be angular, and angularity is the rule in sands of fluvial transportation, particularly in the upper reaches of streams. Russell found the sands of the lower reaches of the Mississippi River more angular than those of the upper reaches. Residual sands generally have irregular shapes and are marked by corrosion pits. These generalizations are based on the assumption that the sands are of original production and are not derived from sands of earlier

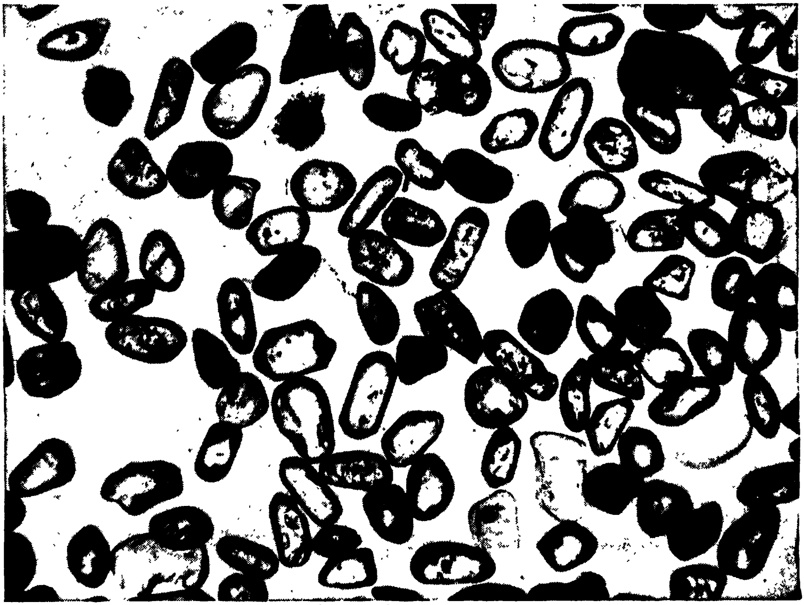

FIG. 35. Nonmagnetic part of $\frac{1}{16}$ to $\frac{1}{8}$ mm. fraction of the same sand as Fig. 33 (× 64). The ellipsoidally well-rounded grains of high relief are zircon, of which a few show the original crystal form. There are a few light-colored grains of quartz. The well-rounded dark grains are dark garnet. Inclusions are present in some grains of zircon. Photographed in liquid of index 1.63.

generation. Rounding of sand grains is obviously a function of the distance of transportation, and long-traveled sands are likely to be better rounded than those that have gone only a short distance. Distance, however, does not mean a long way to the sources, as the travel may have been back and forth on a beach.

A deposit of sands is likely to contain sand particles of many different histories. Thus, glacial outwash may contain sand derived directly from Pre-Cambrian terranes, sand derived indirectly from Pre-Cambrian terranes but directly from some terrane of the Cenozoic, which may have derived it from a Mesozoic terrane, whence it may have come from a Paleozoic terrane, which obtained it from Pre-Cambrian rocks. Imagination may construct many possibilities. The association may be compared to passengers on a great liner sailing from a European port. The passengers may make up an association of many people of many characteristics, each person being, perhaps,

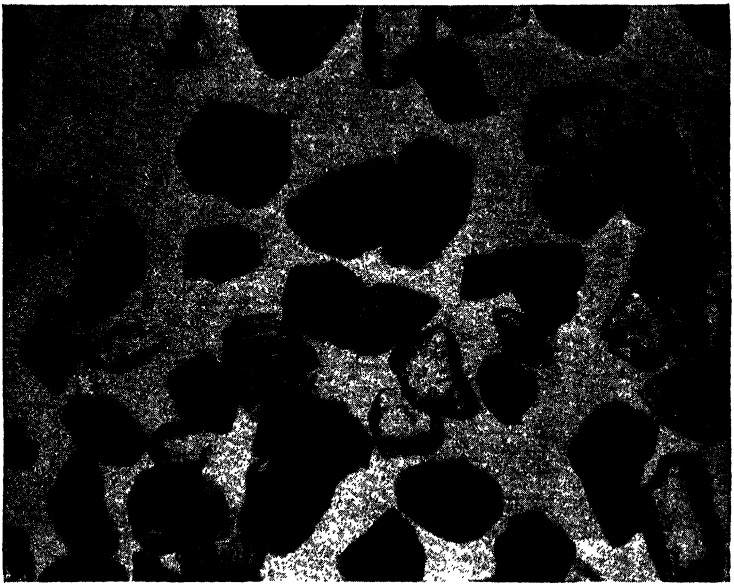

FIG. 36. Beach sands from high-tide level about a mile south of the south jetty at the mouth of Columbia River (× 64). These sands are little rounded. The light-colored grains are quartz, the black grains are ferromagnesian minerals and just right of the center is a green prismatic grain that is euhedral. Photographed in liquid of index 1.63.

from a different country and each having a history different from the others. This fact with respect to sand associations should be realized. A sand of aeolian production may have aqueous deposition; a sand of aqueous production may have aeolian deposition (Udden).

The coasts of Oregon and Washington are bordered by many

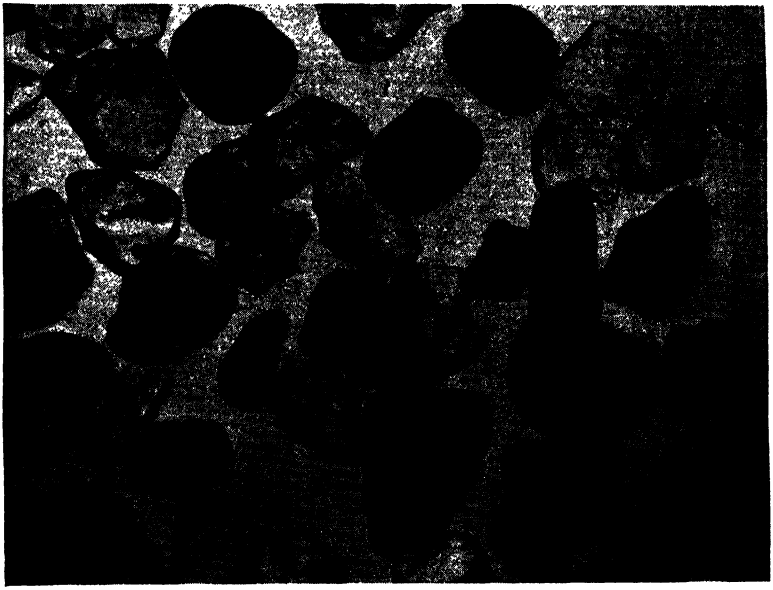

FIG. 37. Dune sand from dune on the Oregon Coast between Saunders Lake and the sea, collected about 1 mile inland from the beach. The light-colored grains are mostly quartz with an occasional grain of feldspar. The dark grains are pieces of rock. Few grains are well-rounded. This sand is typical of the dunes from Cape Falcon south. North of Cape Folcan the sands are angular. The immediate parent of these sands is the beach sand. (× 64). Photographed in liquid of index 1.63.

dunes, and the beaches for miles and miles are of sands. It seems likely that these sands have been in movement on the beaches and in the dunes for millions of years. The sands move from beaches to dunes and from dunes to beaches. Each grain probably has been many times deposited by water and as many times by wind, and it is considered that the sands have come from many sources. Some were derived directly from Tertiary sandstones in coastal cliffs, some were brought by rivers originating in the Coast Range, others from far distant sources as those brought by

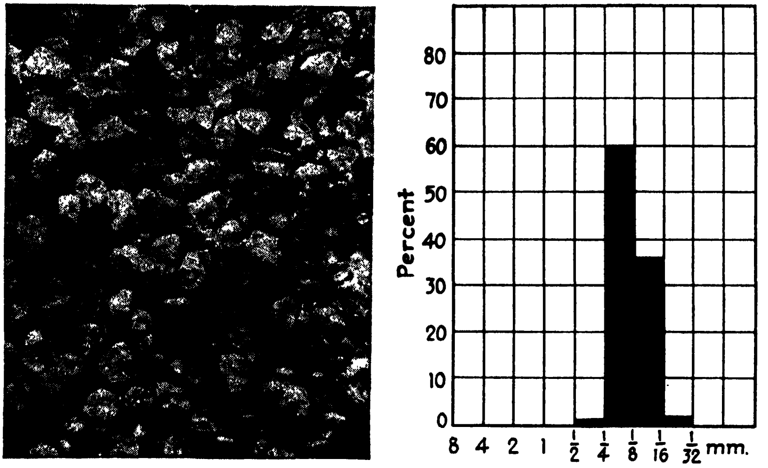

FIG. 38. Microphotograph and histogram of sand from the Bethel sandstone, a formation in the Mississippian system, of marine deposition, probably in connection with a delta. Collected at Cedar Bluff, Caldwell County, Kentucky (× 50). Grains are little rounded in spite of probable long transportation. Grains were probably largely transported in suspension. Sorting is excellent. (*Photograph by H. H. Rasche.*)

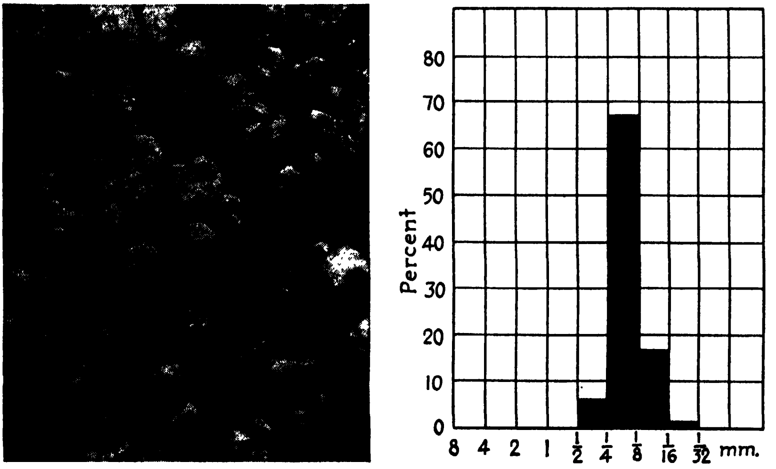

FIG. 39. Microphotograph and histogram of sands from the Pottsville sandstone, Pennsylvanian. Collected from the top of Natural Bridge, Powell County, Kentucky. The grains are small, little rounded, and sorting is excellent. The environment of deposition is considered to have been that of a stream channel. [*Photograph by H. H. Rasche* (× 50).]

the Columbia from Canada. Few grains composed of quartz are well rounded (Twenhofel, 1946).

The sorting of sands ranges through wide limits. Sands of aeolian deposition as a general rule exhibit the best sorting; beach sands take second place; and sands of other environments have greater or less poorness of sorting. However, many sands of beach deposition are as well sorted as any of aeolian deposition, and sands of both aeolian and beach deposition may exhibit poor sorting. Fluvioglacial sands are not unlike those of streams and range from excellently to poorly sorted. There is no sorting in residual sands unless they are the residuum from sandstone. The sands may then approximate the sorting of the parent. It is believed that the degree of sorting is not a reliable criterion of the environment of deposition and that excellently as well as poorly sorted sand may be deposited in any environment of deposition (Figs. 38–40).

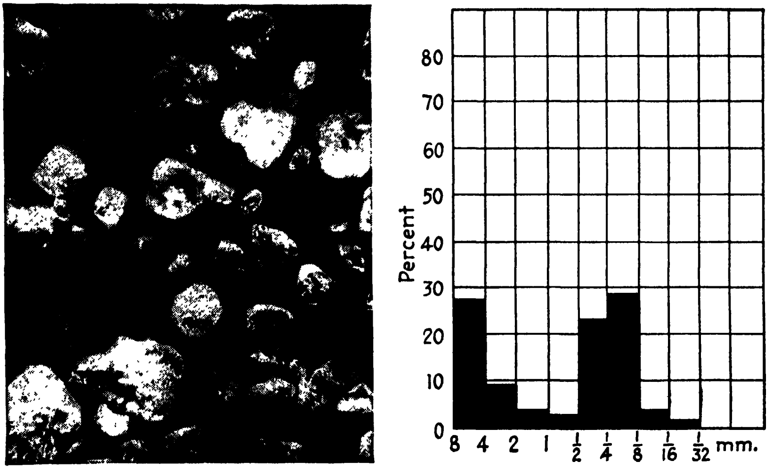

FIG. 40. Microphotograph and histogram of sands from a channel sandstone in the Pottsville (Pennsylvanian) at Diamond Springs, Logan County. The sands were deposited by a stream, and the deposit is a conglomerate in places. The sorting is very poor; the largest grains show a little rounding. (*Photograph by H. H. Rasche.*)

Determination of sorting is difficult, as it is difficult to obtain a sample of a single competency, since the agent of deposition rapidly varies in its competency from time to time and from place to place, so that a sample collected from any place usually represents competency not of a single time, but of several times.

The stratification structures of sands show great variation. Sands of marine deposition tend to have more uniform and continuous bedding than sands of fluvial deposition, and they usually do not vary in character and dimension like the sands deposited in rivers. This is not true for the sands of beaches, which may have extremely irregular depositional structures and sorting. Marine sands may spread over immense areas with more or less uniform stratification. Aeolian sands tend to have irregular stratification. Fluvial sands usually occur as lenses interbedded with gravel, silt, and clay in rapidly changing sequences. They show a cut-and-fill, or flow-and-plunge, type of deposition.

Sands may be cross-laminated in any environment. Cross-laminated units of marine deposition have extensive distribution and are bounded by planes that approach parallelism. The laminations have inclinations in various directions. Fluvial sands also are bounded by nearly parallel planes, but the units do not have extensive areal distribution, and each has a more or less lens shape. Inclinations are in the directions of stream flow. Aeolian sands are bounded by inclined intersecting planes and the cross-laminated units are wedge-shaped. Beach sands tend to be cross-laminated in units of complex pattern but on a small scale. Bounding planes may vary in inclination from horizontal to around 30 deg.

The chief cementing substances of sands are silicon dioxide, iron oxide or hydroxides, and calcium and magnesium carbonates. Iron sulphide or iron carbonate serves as a cement in rare instances. It is not known which cement is most common. The sandstones of the upper Mississippi Valley Cambrian seem to have lime carbonate cement more commonly than any other. Silicon dioxide is frequently deposited in optical continuity with the crystal structure of particles, and the thus-enlarged particles have crystal outlines. Outlines of the original particles often are indicated by slight discoloration. Iron oxide cements may be hematite, göthite, or one of the other hydrous oxides of iron.

Grit is an old name much in use by British geologists as a term applicable to coarse sandstones. It has also been used for sandstones composed of angular grains. It is thought that the name should be abandoned.

Different varieties of sands and sandstones may be indicated by qualifying adjectives. Thus, sandstones containing significant quantities of garnet may be termed garnetiferous sandstones; those containing glauconite, glauconitic sandstones; etc. However, glauconitic sands and sandstones have long been known as greensands, and it is thought that this term will continue to be used, although it is objectionable, as sands other than those containing glauconite may be green. Glauconitic sandstones are very commonly poorly cemented and friable.

The area of distribution of a sand formation varies with the environment of deposition (Dapples). Sands of stream distribution are lenses of variable thickness and limited widths and lengths; those of aqueoglacial deposition may cover large areas between recessional moraines and have semi-fanshaped distribution in front of end moraines; desert sands may cover small to immense areas in the forms of dunes; and marine sands may have lenticular distribution adjacent to some coasts, have narrowly to broadly linear distribution parallel to other coasts, and, if conditions are those of a rising sea level, the sands may be spread as an extensive sheet over much of the surface invaded by the sea with the sands at the place of farthest advance considerably later in time than those at the beginning of the advance. It is believed that the sands of the St. Peter formation of the upper Mississippi Valley represent deposits made by an advancing sea. Cailleux expressed the view that marine and river sands could be differentiated on the basis of the form and appearance of the quartz particles. The writer doubts that this is possible.

Arkose

Arkose (Barton) is a form of sandstone that contains 25 per cent or more of feldspar. The materials of arkose have been derived from the destruction without serious decomposition of acid igneous rocks of granular texture. Granite and granite gneiss are the parent rocks of most arkoses. The feldspars are usually more or less hydrated and are generally little rounded. Sandstones, sands, gravels, and conglomerates that contain considerable feldspar are often described as arkosic.

Problems connected with the origin of arkose relate to the processes and environments that permit granular acid and intermediate igneous rocks, or their metamorphic equivalents, to disintegrate and partially decompose without complete destruction of the feldspars, and to removal and deposition of the released products. These things take place where there is an absence of, or a restricted, vegetation and an adequate slope to permit relatively rapid removal of the products released by weathering. The sites of deposition cannot be far distant from the areas of production.

The conditions for the production of arkosic sediments are found in desert and semidesert areas over which plants maintain no effective cover, in regions of high latitudes and high altitudes where the temperatures are too low to permit adequate protective growth of plants, on steep slopes over which rainfall is so great that a soil cover cannot be established (Krynine), and on coasts and in streams where the force of waves and currents precludes protection. This gives six environments in which arkosic sediments form and near which they are likely to be deposited. The environmental conditions may be placed in the three classes:

1. Rigorous temperature conditions.
2. Moderate to warm temperature conditions.
3. Shores of seas and lakes and channels of rivers.

Arkoses Due to Rigorous Temperature Conditions. These conditions are found in arid and semiarid regions and in high latitudes and altitudes.

Arkosic sediments form in deserts wherever an acid crystalline terrane is present. In order to insure a deposit there should be considerable relief at the site of production to permit removal of the products about as rapidly as released. Places of deposition may be over desert plains, in desert lakes and streams, in humid areas over floodplains, in lakes, and in the sea. The arkosic deposits attain thickness and extent related to the relief of the area supplying the sediments and the position of the base level of deposition of the area receiving them. The sediments have decompositional features proportional to the conditions over the sites of erosion and deposition. They vary in texture in the deposits made in basins from bed to bed and laterally may range from silts at one extreme to gravels at the other. Bedding tends to be irregular to regular, the former a consequence of torrential waters. Cross-lamination may be common, and there may be much mud cracking. There may be much cut-and-fill structure. Colors in desert areas tend to be those of the minerals, with perhaps an intensification due to oxidation. Colors of deposits made in humid areas depend on rates of deposition and quantity of vegetation incorporated with the sediments. The last conditions reduction and, to some extent, bleaching. If deposition is so rapid as to prevent growth of plants, the sediments may have colors like those of desert basins. Some sediments of aeolian deposition may be present, with sands having mat surfaces and aeolian cross-lamination. The sediments are likely to show much lateral variation and to pass in sourceward directions into coarse clastics.

Arkosic sediments of desert derivation deposited in lakes and the sea are not likely to be greatly different from other deposits of shallow water with respect to bedding, sorting, and other depositional features. The sites of lacustrine and marine deposition, however, are likely to support a considerable growth of organisms whose decomposition would lead to bleaching and loss of red color. Arkoses of this origin are stated to be now forming in the Gulf of California (Barton).

Arkoses referred to deposition in an arid environment are the Torridonian sandstones of the Pre-Cambrian of Scotland, the Sparagmite (Cambrian) sandstones of Sweden, a part of the Lower Old Red sandstone of Scotland, and the Paysaten arkose (Cretaceous) of Washington and Oregon (Barton).

High-latitude arkoses may form as residual deposits of limited thickness and extent, or the products of rock destruction may be transported by streams to alluvial fans, floodplains, deltas, or standing bodies of water, in which they have sorting, stratification, and other depositional features correlated with the agents of deposition and the particular environmental conditions. Colors are either those of the composing minerals or the result of bleaching because of the presence of organic matter in the sediments. Little is known of the extent of arkoses of this environment.

High-altitude arkoses may be deposited in high intermontane basins. Extent is generally limited but may be large. Colors tend to be those of the composing minerals, as oxidation usually is limited. Deposition usually leads to more or less incorporation of organic matter, with some bleaching as a consequence. The tendency is to develop a gray color. Arkoses in the Pennsylvania strata of Rhode Island are referred to development in this environment.

Arkoses Formed in Regions of Moderate to Warm Temperature Conditions. There are two phases of this climatic condition: that of semiaridity and that of high rainfall. Both require considerable relief for the production and deposition of arkosic sediments in any quantity.

The arkoses of semiarid conditions may be deposited on land as alluvial-fan, floodplain, and delta deposits, or they may be deposited beneath the waters of rivers, lakes, and the sea. The arkosic sediments deposited on land have sorting, stratification, and other depositional features characteristic of the particular environment of deposition. There may be much mud cracking of those sediments in which it is possible. The sediments originate on uplands in the region, and they have colors and preservation of feldspars and other minerals determined by the extent of decomposition. Deposition must be so rapid as to prevent establishment of a vegetable cover, or the sites of deposition must be too dry to permit such to grow; otherwise colors present on deposition are lost, and the sediments are bleached to some shade of gray. Barton has referred a considerable number of arkoses of the geologic column to deposition in this environment. Among them are the Triassic Sugar Loaf arkose of the Connecticut River valley; the Triassic Stockton arkose of New York, New Jersey, and Pennsylvania; the Permian Rotliegende of Germany; parts of the Old Red Sandstone of Great Britain; and the Custer and Fountain (Permian) arkoses of Colorado.

Arkoses formed in regions of moderate to warm temperature conditions with high rainfall depend upon rainfall to prevent the formation of soil and a vegetable cover. Under these conditions a feldspathic terrane may contribute arkosic sediments directly to sites of deposition, which may be alluvial fans, floodplains, deltas, and the bottoms of standing bodies of water, in each of which the sediments have depositional features characteristic of the environment. Under many conditions, vegetation would become established over sites of deposition, decomposition of the feldspars under the conditions of warmth and moisture would take place, and the feldspars might disappear. Krynine has described arkoses forming under these conditions at the present time in southern Mexico, where the mean annual temperature is 80°F. and the rainfall is 120 in. or more per annum.

Arkoses Formed on Shores of Seas and Lakes and in Channels of Rivers. Arkoses formed from sediments eroded from the shores of the sea, lakes, and rivers may show little decomposition of the feldspars, and the sediments are deposited on beaches and the immediately adjacent shallow water and in stream channels. These arkoses have the depositional features characteristic of these environments. Colors may be those of the composing minerals but, as the sediments are deposited in water containing organic matter, more or less bleaching is probable. Arkosic sands are now being deposited on beaches and over shallow bottoms where coastal rocks contain much feldspar.

Graywacke

Graywacke (Fay; Wentworth, 1935) is defined as the basic equivalent of arkose, having a large content of little decomposed particles derived from basic igneous rocks or their metamorphosed equivalents. Source rocks might also be graywacke as here defined. There would be a large content of ferromagnesian minerals and soda-lime feldspars. Quartz would commonly be a constituent, and clay and silt would also be commonly present.[1] As the ferromagnesian minerals and soda-lime feldspars are generally less resistant to decomposition than the potash and soda-potash feldspars, it may be assumed that very rigorous conditions would be required to prevent their decomposition before deposition. That this is not necessarily the case is shown by the abundance of olivine sands on the beaches of the Hawaiian Islands, by similar abundance of sands derived from Keweenawan lavas on the beaches of the Keweenaw Peninsula and other places on the shores of Lake Superior, and local abundance on parts of the north shore of the Gulf of St. Lawrence. There are undoubtedly many other places. These sands have been produced by erosion of shores.

Any region underlain by basic igneous rocks, if very dry or cold, or very rugged with steep slopes and such great rainfall that a vegetable and soil cover could be difficultly maintained, would present favorable environmental conditions for the production of the sediments necessary for the formation of a graywacke. As basic igneous rocks now form and during much of geologic time have formed much of the surface of the earth, there is little reason to believe that graywacke should be particularly rare. At present there are the vast Columbia Plateau, the plateaus of Deccan and Abyssinia, and other extensive areas in Ireland and South Africa. All of these can and probably do yield basic sands.

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[1] Graywacke is an old German term to which various meanings have been given, among which is that given above. The term has been used for fine-grained dark sandstones in which considerable argillaceous matter is present and for dirty sandstones that have experienced some degree of metamorphism.

Pettijohn (1943) has lately discussed the term of graywacke at length. He does not agree with the definition given above and gives his own and states that it is rock which has undergone some degree of metamorphism. This is the usage it has had for a long time in the Lake Superior region. The present writer is of the opinion that the term has been used with so many different meanings that the wisest course would be to abandon its use.

PYROCLASTIC SEDIMENTS

The pyroclastic sediments consist of materials ejected by explosive activity from volcanoes. Some of these materials are direct derivatives from liquid melted rock and are termed *essential* by Wentworth and Williams, and some consist of materials termed *accessory,* derived from previously solidified volcanic rocks of origin *consanguineous* with the previously mentioned liquid rock. The particles are known as *blocks* (chiefly, if not wholly, of accessory origin), *bombs, lapilli* (cinders), *sand, ash,* and *dust.* Only essential products are considered here, and the term sand is rejected, as its use in this connection is considered not justified. The lapilli and "bombs" have a great variety of shapes and may be spherical, ellipsoidal, pear-shaped, ropelike, or irregular. The shapes of bombs have been described as turtleback, rotational, spindle, bipolar, unipolar, and tear-shaped. Some lapilli have oölitic and pisolitic internal structure. These are termed *accretionary lapilli* by Wentworth and Williams and range in dimension from 2 to 10 mm. They are said to form through accretion of fine "ash" or dust around raindrops falling through an explosion cloud or are accretionary to a nucleus rolling on the ground. Pele's tears are small glass drops or pairs of drops in dumbbell arrangement which form in some eruptions. They range to a few millimeters long. *Pele's hair* is a natural lava wool that may be found in any pyroclastic sediment. "Ash" particles are composed of fragments of glass that represent walls between bubbles in lava froth or foam.

Blocks may be several feet in diameter, and spheroidal "bombs" up to 5 ft. in diameter are not uncommon. Ropelike bombs attain lengths to 10 ft or more.

Classification of pyroclastics has been made by several students, among whom are La Croix and T. B. Bailey. The classification used in this work is that of Wentworth and Williams. Each grade is composed of 50 per cent or more of the dimensions indicated. With some modification the classification is as follows:

TABLE 18

Dimensions, mm.	Fragments	Lithified products
256 or greater	Large "bombs"	Very coarse-grained volcanic breccia or agglomerate
32 to 256	Large lapilli	Coarse-grained volcanic breccia
4 to 32	Small to medium lapilli	Fine-grained volcanic breccia
¼ to 4	Coarse "ash"	Coarse-grained tuff
¼ or less	Fine "ash" and dust	Fine-grained tuff

The coarse-grained pyroclastics fall near places of expulsion as dimensions and conditions dictate. Stratification and sorting are very poor or wanting. Fine-grained pyroclastics are classified in their descent, and sorting and stratification may be excellent. They may be transported many miles from sources by wind.

The chemical and mineralogical composition of pyroclastics varies with the composition of the lava of derivation. Most are composed of glass.

Pyroclastics on deposition mantle the irregularities of the surface, forming domes, inverted domes, anticlines, synclines, and terraces, and, if water does not modify deposition, the thickness on the elevations tends to equal that in the depressions.

The fine-grained pyroclastics may fall in large quantities on water and may also be picked up by water subsequent to deposition. They then become parts of other sedimentary deposits and may lose identity.

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THE FINE-GRAINED CLASTICS

INTRODUCTION

The fine-grained clastics include those particles in the range from $1/16$ mm. downward. There are two classes with the separating dimension at $1/256$ mm. or about 4 μ. The coarser grade is designated silt, the finer clay (Correns). Other classifications have been made. Most, if not all, silts contain much material both below and above the limiting dimensions, and also most, if not all, clays contain more or less silt and sand.

The indurated equivalents of clays are claystones (Terzaghi, 1937), and the silts when indurated form siltstones. Clays and silts mingled with water form muds, and both claystones and siltstones have been termed mudstones. If claystones and siltstones have cleavage parallel to bedding, the term shale should be applied. This cleavage is either primary, and due to deposition, or secondary, due to readjustment of colloidal particles present in most silts and all clays, or it may be due to development of authigenic minerals with alignment parallel to bedding.

The change from clay to claystone and silt to siltstone is connected with elimination of water, compression of particles so as to place them closer together, recrystallization, and enlargement of particles. The changes probably begin on deposition and continue for an extremely long time. The induration is indicated as a rule by claystones and siltstones retaining considerable coherence on being wetted after drying, whereas dry clays and silts not containing bituminous matter on being wetted disintegrate into discrete particles. Naturally there is no sharp division, and many formations termed claystones and siltstones contain parts to which these terms do not apply. The diagram (Fig. 41) shows the essential features of the classification based on induration.

Clays and silts are composed of clay minerals, iron oxides and hydroxides, silicon dioxide, and fine particles of common and many uncommon minerals. The composing materials are thought to be dominantly of decompositional origin. There is much material of colloidal dimension (Ashley). Silts are composed of small particles of many kinds of minerals, of which many have been produced by rock abrasion, grinding, and impact, and others by decomposition. Particles of physical production are ordinarily thought to compose the largest quantity. Allogenic quartz, garnet, feldspar, mica, and many other minerals are known to be present in clays and silts. There is also much water which is held in subcapillary spaces and behaves as if solid. Many clays are clastic which is referred to the presence of small scaly particles (Terzaghi). There are all gradations from clays to silts, and most clays seem to be mixtures of particles of the two grade sizes.

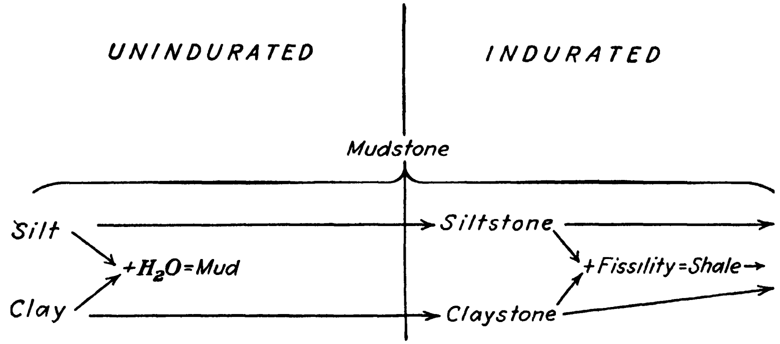


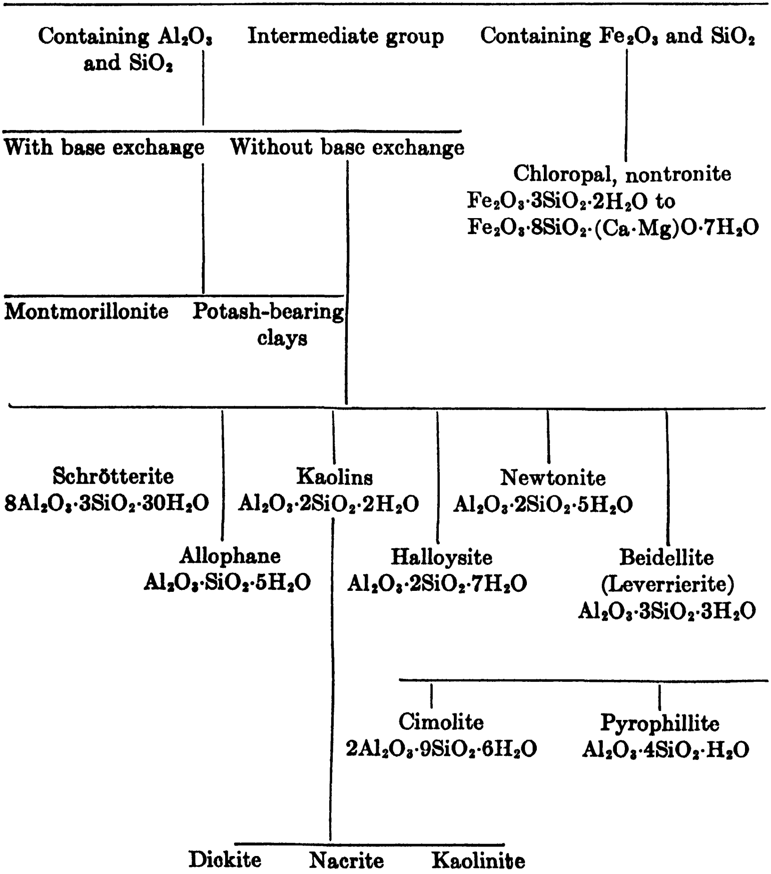
FIG. 41. Diagram showing relations of fine-grained clastic sediments. The arrows on the right indicate that induration continues to form argillites and slates.

The clay minerals have flakelike shapes and are hydrous aluminum silicates. According to Ross and Kerr they may be placed in four groups as follows: (*a*) kaolinite, (*b*) montmorillonite-beidellite, (*c*) potash-bearing clays, and (*d*) a group of which the properties have not been definitely determined.

The kaolinite group contains kaolinite, dickite, nacrite, halloysite, and endellite (Ross and Hendricks). Kaolinite is probably the most important of this group. It is common in residual clays and is the clay mineral most commonly found where organic acids have subjected the products of decomposition to prolonged leaching. Nacrite and dickite are generally rare, and hypogene alteration seems to be responsible for most occurrences. Halloysite has wide distribution and is interpreted by Ross and Kerr as a form of kaolinite with the mineral particles in a fine or submicroscopic state of division. The composition of these minerals is shown in the table of Marshall, Table 19.

TABLE 19. GRAPH SHOWING THE CLAY MINERALS[1]

The Clay Minerals


Containing Al_2O_3 and SiO_2
Intermediate group
Containing Fe_2O_3 and SiO_2
With base exchange
Without base exchange
Chloropal, nontronite
$Fe_2O_3 \cdot 3SiO_2 \cdot 2H_2O$ to
$Fe_2O_3 \cdot 8SiO_2 \cdot (Ca \cdot Mg)O \cdot 7H_2O$
Montmorillonite
Potash-bearing clays
Schrötterite
$8Al_2O_3 \cdot 3SiO_2 \cdot 30H_2O$
Kaolins
$Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$
Newtonite
$Al_2O_3 \cdot 2SiO_2 \cdot 5H_2O$
Allophane
$Al_2O_3 \cdot SiO_2 \cdot 5H_2O$
Halloysite
$Al_2O_3 \cdot 2SiO_2 \cdot 7H_2O$
Beidellite
(Leverrierite)
$Al_2O_3 \cdot 3SiO_2 \cdot 3H_2O$
Cimolite
$2Al_2O_3 \cdot 9SiO_2 \cdot 6H_2O$
Pyrophillite
$Al_2O_3 \cdot 4SiO_2 \cdot H_2O$
Dickite
Nacrite
Kaolinite

[1] After C. E. Marshall, Clays as minerals and colloids, *Trans. Ceramic Soc.*, 30 (1931), 81–97.

The montmorillonite-beidellite group has wide distribution. Ross and Kerr give five members of which one is not named but has the composition of kaolinite. The other four are beidellite, nontronite, montmorillonite, and saponite. Ross and Hendricks list, in addition, hectorite and valchonskoite. Beidellite commonly contains ferric iron. It is common in many marine clays and claystones, metabentonite, soils of the cooler and moist sections of the United States, and soils of sections where leaching is not extensive. Clays and silts may contain several clay minerals which may be referred to alteration of some of them or to deposition of clay minerals derived from different provenances. Montmorillonite is the clay mineral of some fuller's earth, the Cretaceous bentonites of the Great Plains area, and it is also a constituent of clays derived from the decomposition of various minerals of igneous rocks.

The potash-bearing clays are those characteristic of the Ordovician metabentonites of the upper Mississippi Valley, Ontario, and the Appalachian region.

Marshall classifies the clay minerals first on the basis of composition and second on the basis of base-exchange data. His classification with some modification is given in Table 19.

According to Grim (1939, page 472), the base exchange capacities of montmorillonite, illite, and kaolinite are 60–100, 20–40, and 3–15, respectively.

The extreme fineness of the particles of clays and silts gives tremendous surface area, and it has been estimated that a gram of clay may have 200 to 900 square meters of surface. Such extensive surface gives extremely great adsorption capacity to clay, and certain substances are tenaciously held. Prolonged leaching of clays and silts removes the salts of sodium, magnesium, and calcium, particularly if the leaching is due to water containing organic acids. Potash tends to remain with the clays, explained by some as due to the development of secondary minerals, of which one, generally termed sericite, and lately named illite (previously named bravaisite by Mallard in 1878) (Grim, Bray and Bradley; Correns) is considered the most common, and by others as due to adsorption. Fine-grained unleached clays may approximate in quantity the calcium, magnesium, sodium, and potassium in the parent rocks. In leached clays the potassium-sodium ratio is higher than in the parent rocks and, in general, higher than in the unleached clays. Clays when introduced into waters other than those in which they originated may change in content of sodium, potassium, calcium, and magnesium, owing to base exchange. The alumina content of leached clays also seems to be higher than those not subjected to much leaching. This may be due to removal of silica not combined with clay minerals and in some cases to decomposition of clay minerals followed by removal of silica.

According to Dietz the formation of montmorillonite is favored by an alkaline environment and absence of leaching, presence of ferromagnesian minerals in parent materials, and presence of bases. If there is a strong concentration of potassium in the surroundings, the potassium clays form instead of montmorillonite. Ross and Hendricks state, "In general, alkalic feldspars and the micas tend to alter to kaolin minerals." The parent rocks of minerals of the montmorillonite group are those composed of ferromagnesian minerals, calcic feldspars, and volcanic glasses. Formation of minerals of the montmorillonite group is favored by alkaline conditions, acid conditions favor formation of minerals of the kaolinite group. Presence of magnesian ions in the altering system favor formation of minerals of the montmorillonite group. Climatic factors modify results. Conditions unfavorable for prolonged decomposition with potash available, as would be the case in marine waters, favor formation of the bravaisite-illite group. Prolonged leaching under high temperature or tropical conditions with ferric iron in parent rocks favors formation of minerals of the kaolinite group.

The particles of clays and silts are commonly transported in suspension, and the particles are so small that they are effectively cushioned in both air and water against abrasion, so that shapes are largely those of solution, decomposition, and fracture.

Deposition takes place after the competencies of the transporting agents have decreased to low value. Clays and silts tend to be separated from the coarse and medium clastics with which once they may have been transported in the same currents. They may be deposited above coarse clastics of preceding currents and become overlaid by coarse clastics of succeeding currents.

Classification of clays and silts and their indurated equivalents on almost any basis gives little satisfaction to other than the classifier and often not even to him. Color will not serve except superficially. Classification on the basis of composition is difficult and often useless and in most cases requires chemical analysis. Clays have been classified as gypsiferous, argillaceous, siliceous, arenaceous, ferruginous, glauconitic, carbonaceous, bituminous, etc. These terms in some cases denote a definite characteristic, as for instance glauconitic, but the term ferruginous is applied only when the fine-grained sediment is red, although it is well known that the quantity of iron may be very small and that many nonred sediments contain much more iron than those that are red.

TABLE 20. ANALYSES OF CLAYS AND SILTS

Compound	Residual clay,[1] per cent	Glacial clay,[2] per cent	Mississippi delta composite,[3] per cent	Lake-bottom sediment,[4] per cent	Red muds,[5] per cent	Red clays,[6] per cent	Blue and gray muds,[7] per cent	Green muds,[8] per cent	Bituminous shale,[9] per cent
SiO_2	55.90	48.81	69 96	59.18	31 66	54.48	64.20	31.27	51.03
Al_2O_3	19.92	7.54	10 52	17.30	9.21	15.94	13.55	4.08	13.47
Fe_2O_3	7.30	2.53	3.47	6.80	4.52	8 66	8.38	12.72	8.06
FeO	0 39	0.65				0.84			
MgO	1.18	7.05	1.41	2.31	2.07	3.31	0.61	0.39	1.15
CaO	0.50	11.83	2.17	1.16	25.68	1.96	5.08	27.32	0.78
Na_2O	0.23	0.92	1.51	....	1.63	2.05	....	....	0.41
K_2O	4.79	2.60	2.30	....	1.33	2.85	....	....	3.16
TiO_2	0.20	0.45	0.59	....	......	0.98	....	....	
P_2O_5	0.10	0.13	0.18	....	..... .	0.30	0 64	0.32	0.31
CO_2	0.38	15.47	1.40	....	17.13	....	1.69	20.25	7.29 (S)
H_2O	9.06	2.02	5.74	(Ig.) 8.46	6.02	7.04	5.60	3.30	0.81
C	.....	....	......	....	......	.... .	.. ..	....	13.11
Misc.	.....	0.50	1.25	....	2.73	1.59	0.25 SO_3	0.35 SO_3	3.32 (Hy)[9]
Total	99.95	100.50	100.50	95.21	101.98	100.00	100.00	100.00	102.90

[1] Residual clay from limestones. Analyses by Steiger, *U.S. Geol. Surv.*, *Bull.* 770, 511.

[2] Glacial clay from Milwaukee. Chamberlin and Salisbury, *U.S. Geol. Surv.*, *5th Ann. Rept.*, 578.

[3] Analysis of composite of 253 samples from Mississippi Delta. Analyses by Steiger, *J. Washington Acad. Sci.*, 4 (1914) 59.

[4] Depth of 630 feet in Lake Ontario. Kindle, *Trans. Roy. Soc. Canada*, 19, sec. 4 (1925), 56.

[5] Red muds. Analyses by Hornung, Murray, and Renard, "Deep Sea Deposits," pp. 444–445.

[6] Red clays, composite of 51 samples. Analysis by Steiger, *U.S. Geol. Surv.*, *Bull.* 770, 518.

[7] Blue and gray muds from bottom of Atlantic Ocean. Murray and Renard, "Deep Sea Deposits," p. 197

[8] Green muds. Murray, and Renard, "Deep Sea Deposits," p. 449.

[9] Bituminous shale from Dry Gap, Ga. *U.S. Geol. Surv.*, *Bull.* 770, 552. Analyses by Eakin; 2.73 per cent should be subtracted from the total, for oxygen added in the analysis. Hy indicates hydrocarbons; S, sulphur.

This work attempts a classification on the basis of the environments and processes responsible for deposition. It is little more satisfactory than a classification based on color or composition, but it is convenient. The sediments are placed in the seven classes of residual, glacial, fluvial, lacustrine, marine, aeolian, and volcanic. Fine-grained sediments of any color, composition, and texture may form in most of these environments. Table 20 shows chemical analyses of clays from each of the environments. The analyses are thought to be representative, but it should be appreciated that the clays and silts from each environment exhibit much variation in composition. It should be noticed that the silica content in each analysis is more than is required to account for the clay minerals. It is thought that this silica is fine-grained quartz and/or colloidal silica deposited at the same time as the associated and fine-grained sediments.

RESIDUAL FINE-GRAINED CLASTICS

Residual fine-grained clastics have undergone no transportation other than creep and hence are unstratified and unsorted. They commonly contain large residual particles of various shapes and dimensions. Colors range from white to red, the lighter colors being commonly found in the colder parts of the temperate zones. Red colors are found in warm to tropical regions of good drainage. The red of lateritic soils is the extreme development of this color. Intermediate regions have yellow clays and silts. Black clays and silts are found in steppe regions, as the regur of India and the tschernosem of Russia. Arid and semiarid regions tend to have light-colored clays and silts in the temperate and subtropical latitudes, but they may have red clays and silts as are some in the Sahara. Residual clays and silts of humid regions are generally low in calcium, magnesium, and sodium salts, but those of arid to semiarid regions are likely to have a high content of these salts.

Under certain conditions in tropical and subtropical regions, clay minerals decompose to aluminum hydroxide and silica, and iron minerals are reduced to iron hydroxide. The silica is removed and the resulting residual product is *laterite*, which at one extreme is composed of iron hydroxide and at the other of aluminum hydroxide. In cool regions, iron is reduced to the ferrous condition and removed in solution, leaving the clays and silts composed of hydrous aluminum silicate and silica. This is the *podsol.*

Residual fine-grained clastics mantle existing land surfaces. They are not common in the geologic column.

Glacial Fine-grained Clastics

The fine-grained sediments produced by glacial action are largely the result of grinding, abrasion, and impact. They are essentially rock flour. Fully decomposed materials in considerable part may be found in the terminal moraine, but there is little elsewhere unless it was produced after deposition. Fine-grained sediments of ice deposition are unsorted and unstratified. Aqueoglacial clays and silts have various degrees of sorting and stratification. As powdered rock is generally light colored, it results that glacial clays and silts are generally light colored. Exceptions are the rock flours derived from red rocks.

The composition of glacial clays and silts depends upon the rocks that are reduced to fine particles, and hence there may be great variation in composition. There may be no clay minerals. The content of carbonates may be high and the potash-sodium ratio low to high. They contain little or no organic matter. An analysis is given in Table 20.

Clays and silts of ice deposition are mingled with larger particles of which some may have weights of tons. Those of melt-water deposition tend to have large particles wanting except as they are dropped from floating ice. They may be interbedded or dovetail with sands and gravels.

Fluvial Fine-grained Clastics

The fine-grained clastics of fluvial deposition tend to be more or less poorly sorted and irregularly stratified. There is much filling of contemporaneously eroded depressions and interbedding with sands and gravels. They tend to have irregular distribution over both deeps and shallows of the bottom. There are rapid variations from silts and clays to sands in both the vertical and horizontal distribution.

Fluvial clays and silts of humid regions tend to be lower in soluble matter than those of arid and semiarid regions, in which there may be a high carbonate and sulphate content. Organic matter is likely to be abundant in the clays and silts of humid floodplains, and it may be equally so in the clays and silts of arid and semiarid floodplains.

Fluvial clays and silts of humid regions tend to be dark to black at the surface and to become blue to gray with depth.

Clays and silts of semiarid to arid floodplains may have brown to red colors, particularly if they were derived from moist and warm well-drained uplands, and they may also contain concretions of calcium carbonate and sulphate and, more rarely, iron oxide. Cold regions have fluvial clays and silts with colors ranging from light to black. An analysis is given in Table 20.

LACUSTRINE FINE-GRAINED CLASTICS

The fine-grained clastics of lacustrine origin tend to be well sorted, and most of them are extremely fine-grained. Bedding may be in very thin units unless the deposits are subject to much reworking by organisms in which case detailed stratification is not likely to be present. Lake sediments ordinarily contain much organic matter of which some grew in the lakes and some was received via streams and the atmosphere. This accumulates with the inorganic sediments and, if not eliminated by organisms, gives black color to the sediments. Sediments of shallow lakes in regions yielding hard water usually contain much calcium carbonate. This is not present in lakes containing soft water, and it also may be absent over deep bottoms of hard-water lakes (Kindle). An analysis of lacustrine clay is given in Table 20 and a profile of the sediments of a hard-water lake in Fig. 15.

MARINE FINE-GRAINED CLASTICS

Fine-grained clastics of marine origin have such wide variations that few generalizations have application. However, they tend to be well sorted and may be well stratified. Colors are various, and composition depends on many variables. Most have a considerable carbonate content. They may be placed in two classes, of which one is about the continental margins and the other on deep bottoms far from land. The sediments of the former are derived more or less directly from the land and have composition influenced by the regions of derivation. These sediments were precipitated either by loss of velocity of fluvial or marine currents or by flocculation and settling. Colors are red, gray, blue, black, green, or some blending of these colors. Clays of the second class received contributions directly from the atmosphere; directly from the waters that transported them from the land; indirectly from organisms in insoluble parts of shells, bones, and plant matter; and directly from volcanic sources. Different kinds of the marine fine-grained clastics are considered in the paragraphs that follow.

"Red Muds." The term "Red Muds" is applied to clays and silts that at present are in process of deposition over several areas of the ocean bottom having aggregate area of about 100,000 square miles. These muds are known to extend to depths in the oceans of about 2,000 meters. The largest areas are about the mouths of the Amazon, Orinoco, and Yangtze rivers.

The color of these muds is reddish brown, composition is not greatly different from that of other marine shallow-water clays and silts, and the lime content varies with locality from 6 to 60 per cent. Organic matter consists of sponge spicules, diatoms, and other shells to the extent of about 30 per cent. The iron content is low. The red color seems to be largely confined to the surface and to disappear with depth, presumably owing to reduction of iron by organic matter. An analysis of red muds is given in Table 20.

Blue and Gray Muds. The blue and gray muds of the ocean bottom range in depth from just below sea level to depths slightly exceeding 5,000 meters, and they cover about 4,000,000 square miles of the Atlantic, 3,000,000 square miles of the Pacific, 1,500,000 square miles of the Indian, 1,500,000 square miles of the Southern, and 2,500,000 square miles of Antarctic bottoms. The muds when dry tend to lose the blue color and to become gray. The composing materials are of both organic and inorganic origin, the former to about 15 per cent. Colors range from reddish to brownish at the surface, but beneath the surface the colors of the wet muds are gray to blue. Hydrogen sulphide is rather constantly present.

The physical constituents of the gray and blue muds vary with locality and nearness and character of the adjacent lands. Particles of terrigenous origin usually are small and therefore angular. The upper range in dimension is about 0.03 mm., and the range in quantity of terrigenous materials is to 75 per cent. The lime content ranges from zero to about 35 per cent, with the smallest lime content, in general, in the greatest depths. Glauconite may or may not be present. The conditions seem to be too reducing for its development. A chemical analysis of gray and blue mud is given in Table 20.

Green Muds. Green muds are thought to owe their color very largely, if not entirely, to finely divided glauconite, and they are commonly found more or less associated with glauconitic sands. They have been estimated to cover the ocean bottoms to the extent of about 1,000,000 square miles in depths of less than 100 to over 2,500 meters. They tend to lie near the edge of the continental shelf. The physical composition of the green muds, aside from the presence of glauconite, is not unlike that of the gray and blue muds except that there does not seem to be so much organic matter. Small to large nodules of calcium phosphate may be present, and some of the glauconite is of sand dimensions. Calcium carbonate is generally present in the range to about 50 per cent. A chemical analysis of green mud is given in Table 20.

Red Clays. Red clays are distinguished from the red muds on the basis that they are deposited on deep bottoms far from land and have been oxidized by sinking of the composing materials downward from the surface through thousands of feet of water, whereas the red muds are deposited in comparatively shallow water not far from land, and the sediments are largely in an oxidized condition when they reach the sea. The color ranges from brick red in the North Atlantic to a dark chocolate color in the South Pacific. The red clays are extremely fine-grained, over 85 per cent being of fine silt or clay dimension, plastic, and seemingly homogeneous. They may contain resistant parts of many organisms, as ear bones of whales and shark teeth, tests of protozoa and diatoms, and shells of deep-water benthos. Inorganic particles of macroscopic dimensions consist of various rafted rock fragments, pyroclastics, and pumice. Cosmic particles are occasional. Particles, possibly resulting from diagenesis, are nodules of manganese and calcium phosphate and zeolites such as phillipsite. Calcium carbonate is commonly present in the range from a trace to nearly 30 per cent. The content decreases with increase in depth of bottom. Except for rafted materials, allogenic mineral particles are very small and in the range from 0.1 to 0.85 mm. Most of the red clay seems to be below the dimensions of fine silt.

The red clays form one of the most extensive of marine deposits, and they cover an estimated area of 51,500,000 square miles. The Pacific Ocean has over 40,000,000 square miles, the Atlantic Ocean nearly 6,000,000 square miles, and the Indian Ocean nearly 5,000,000 square miles. The range in depth is from about 2,500 meters to the deepest bottoms. An analysis is given in Table 20.

Black Muds (Conacher, Goldman). Black muds of marine origin acquire their color from black sulphides of iron and organic matter. Blackness due to black sulphides of iron is lost as these sulphides change to marcasite or pyrite. Organic matter may consist of humic materials low in fatty substances or bituminous materials high in fats, waxes, and gums. Almost invariably associated with the accumulation of organic matter in salt waters are the black sulphides of iron, and black muds of marine deposition almost invariably contain them. An analysis of black shale is given in Table 20.

Marine black muds form in lagoons, sounds, and bays of bodies of water with weak tides, such as the Baltic Sea (Twenhofel, 1915), and in deep holes on neritic bottoms. Extensive areas of almost unpopulated black-mud bottoms have been recently described in the Arabian Sea (Lees). They do not seem to form on elevated or level parts of the bottom of the open sea; conditions for formation seem to be closely related to no (or limited) circulation. Depth and temperature do not seem to be essential factors, but both are of secondary importance, as low temperature tends to retard bacterial activity and depth tends to decrease circulation. The waters are unfit for homes of many marine animals because of the hydrogen sulphide and sulphuric acid in the water and the absence of oxygen. Scavenger work is thus limited, and organic matter is not eliminated. Blackness of color thus persists after the black sulphides of iron become marcasite or pyrite.

Black-mud bottoms support few benthonic organisms. Plankton and nekton may dwell in immediately overlying waters, but they do so at their peril, as the hydrogen sulphide and sulphuric acid produced in the muds may lead to wholesale killing. Currents from populated waters may bring an abundance of life to the areas with black-mud bottoms, to meet death as if destroyed by a pestilence. The dead bodies settle to the bottom where they are not devoured by scavengers, as none dwell there, and thus the most delicate structures may be preserved.

Black shales derived from black muds rich in organic matter are not infrequent in the geologic column. Among them are the Chattanooga, Utica, Deepkill, and Normanskill black shales and the black shales associated with the "Coal Measures." The last are characterized by a high content of carbonaceous matter and ordinarily little pyrite or marcasite. In most cases the "Coal Measures" black shales are of continental origin and were formed on floodplains or in fresh-water lakes.

The marine black shales tend to be evenly bedded and often thinly laminated. Layers of sandstone to 1 ft. in thickness are not uncommon, and layers of conglomerate have been noted. Pyrite or marcasite nodules are of greater or less commonness. These shales contain few large fossils of benthonic habitat, and planktonic and nektonic fossils are usually rare, but some layers contain fossils in abundance. These are usually delicate planktonic forms, such as the graptolites of the Ordovician and Silurian black shales and the young shells of animals that are a part of the plankton in early life but become benthos in mature life, the problematical conodonts, occasional fish remains, and spores and spore cases. Shells of cephalopods and fragile shells of certain trilobites are occasional.

Organisms with chitinous structures are preserved in black shales as films of carbon; those composed of calcium carbonate may be preserved as calcite, but not infrequently the fossils of these organisms are in the form of marcasite or pyrite.

The black shales have been variously interpreted. Clarke and Pompeckj referred certain black shales to deposition in an environment like the Black Sea; Schuchert considered that deposition may have been in places of poor circulation, such as more or less closed arms of the sea and deep holes of the bottom, the last termed *halistas* by Walther and *dead grounds* by Johnstone, and perhaps beneath sargasso seas. Reudemann postulated a deep-water environment with limited circulation, owing to depth. He followed Marr in assuming deposition within a large gulf or bay. Ulrich stated that enclosed and stagnant conditions are not essential for formation of the black shales. Grabau and O'Connell considered that the black shales of southern Scotland and southern Sweden were deposited in deep water, but Grabau later advocated the view that the *normal graptolite shales*—shales consisting of layers containing graptolites interbedded with thick layers of other colored shales and shaly sandstone layers without graptolites—were deposited on floodplains occasionally inundated by the sea, the graptolite layers representing the deposits of the inundations and the layers without graptolites the deposits of the floodplains. This interpretation does not seem tenable. He considers the *black-shale type* to have been deposited in estuaries and over mud flats bordering alluvial plains. Scupin suggested that the black shales studied by him resulted from periodic marine flooding of lowlands protected from the sea by some sort of barrier.

It is Raymond's view that the color of the Lower Paleozoic black shales is due to concentration of chitinous skeletal matter which drifted offshore from shallow water to the open sea where it was deposited with inorganic muds on the outer slopes below the depth of the profile of equilibrium. He also believed that such chitinous skeletal matter was more abundant at that time than later. He even seems to believe that the black graptolitic shales of the early Paleozoic were deposited under aerobic conditions. The author does not agree with this view. Twenhofel suggested that some black shales were shallow-water deposits of inland seas with weak tides that invaded or retreated from lowlands so that there was a wide shallow-water belt covered with aquatic plants which was bathed with fresh water on the landward side and with salt water on the seaward side, with boundary between the two waters a fluctuating one. During some seasons this boundary may have been such that salt waters extended to the shores; during other seasons it may have been near the outer margin of the aquatic plants. This would have created a very unstable condition so that normal marine organisms could not have inhabited the bottom, scavengers would have been rare, and organic matter would not have been destroyed. Black shales would have been deposited, and these would have risen in the geologic column either as the sea invaded or retreated. Woolnough and Strøm postulated that the deposition took place in barred or landlocked basins under conditions that did not permit circulation of bottom waters, which as a consequence were deprived of oxygen, causing the bottoms to have an absence of benthonic life. The barred basins of Woolnough were of the nature of lagoons that developed on a peneplaned surface as it became submerged by rising sea level. The landlocked basins of Strøm are bays and fiords with narrow connections with the sea and an elevated bottom at the threshold so that the basins are deeper within than at the entrance. This is the condition in many of the fiords of Norway (Fig. 42) and Newfoundland. In humid regions the water level in basins of any origin would be maintained by flow of fresh water from the lands, and the waters might be fresh or brackish. In dry regions the level would be maintained by flow from the sea, and the salinity would be above normal. Only storms would bring into the basins of humid regions waters from the sea, and there would thus be only episodic entrance of marine organisms. In dry regions they would come in constantly through entrances as the water level was being maintained from the sea, but at times storm waters might bring in large numbers in a short time. Introduction of the marine

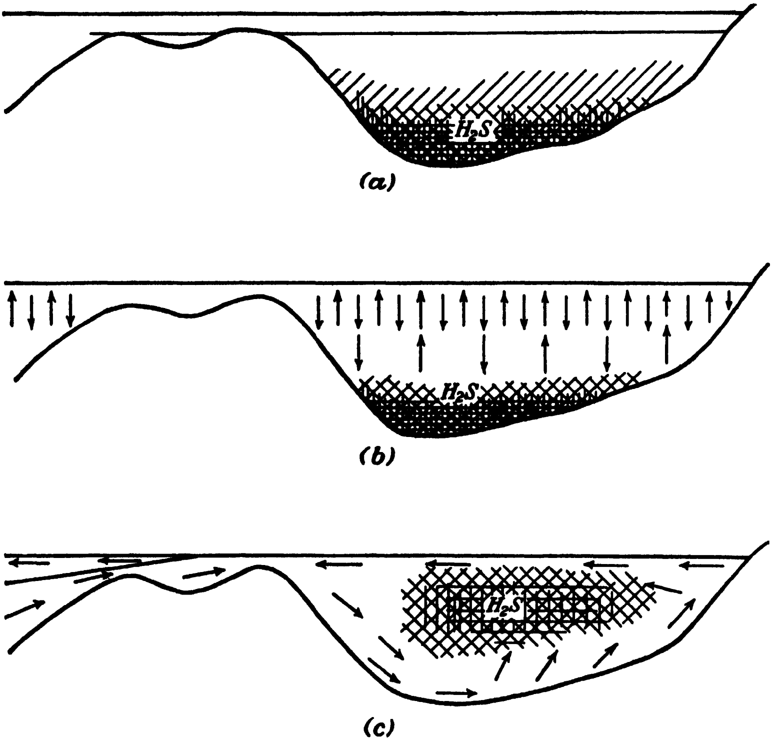


FIG. 42. Section through a typical landlocked basin. (*a*) basin without ventilation of the bottom waters; (*b*) basin with ventilation in the upper part of the bottom waters; (*c*) basin with circulation produced by flowing of heavy salt water from the sea. (*After Strøm, p.* 10.)

organisms into an environment to which they were not adapted would lead to death and ready entombment as the dead bodies settled into the soft muds covering the bottom. It is suggested that the deposits of the basins in humid regions gave rise to the normal black shales; the deposits of dry regions to the black-shale type.

At the present time the oceans have a deep circulation from the poles to the equator beneath and return to the poles above. This provides the bottom waters with oxygen. If the climate of the earth were somewhat equable, this circulation would not exist and it is probable that the bottoms in many places would become low or wanting in oxygen, under which conditions there might be black-mud deposition on the deep ocean bottom over the areas of oxygen deficiency (Strøm).

Thus, it is believed the conditions for black-shale formation are the accumulation of mud and organic matter either in quantities too great and too rapidly for elimination of the organic matter by scavenger and bacterial organisms, probably very rarely the case, or by deposition under conditions in which organic activity leading to elimination of organic matter is prevented. The latter condition requires no (or limited) circulation of the overlying waters so that entrance of oxygen is prevented. Such limited circulation exists in many lakes without through-flowing waters, in tideless seas, in landlocked basins (the barred basins of Woolnough), bottoms of relic seas like the Black Sea, deep holes in the shallow sea bottom, marginal lagoons of various origin, shore waters of shallow inland seas with weak tides and low margining lands, probably over the deep ocean bottom under conditions of equable earth climate, and doubtfully on flat lands subjected to marine inundations, as postulated by Grabau. It is obvious that each occurrence of black shale constitutes a distinct problem and that the environment of deposition of each black-shale formation must be determined on the basis of the associated facts. So far as the writer is acquainted with the black shales of North America, none has been seen that justifies assumption of deep-water deposition.

Fine-grained Clastics of Volcanic Origin

Explosions of volcanoes expel large quantities of materials of clay and silt dimension. Some particles are composed of glass produced by the explosion of gases in lava froth and foam. These are generally known as volcanic "ash." Other particles consist of small fragments detached from the walls of conduits or craters. These materials fall around the volcanoes or are carried short to long distances leeward of the places of origin. They may be deposited without inclusion of materials from other sources, or they may be mingled with other sediments in greater or less quantity. Original colors tend to be light, but the mingling of other sediments may change the color. The particles that fall in water tend to be well stratified; those deposited on the land tend to be unstratified. The particles produced by explosions of lava froth are triangular, as seen in two dimensions, with the sides of the triangles reentrant curves. Particles blown from the sides of conduits or craters may have any shape.

Volcanic "ash" is present in many deposits of the present sea bottom and has extensive distribution in the geologic column. There are large deposits in the Tertiary of western United States. It has been estimated that about 5 cubic miles of volcanic "ash" were ejected in 1912 during the eruption of Katmai in Alaska, and if the "ash" ejections of all time were added, the sum would unquestionably be of stupendous proportions.

Scattered throughout the geologic column are deposits known as bentonite or metabentonite. Bentonite has wide distribution in the Cretaceous of the western part of the United States, and it has been identified in the Tertiary of Argentina, Poland, and North Africa. Pure and unaltered bentonite has white, blue, pale-green, or cream color and a somewhat waxy luster. It has ability to absorb as much as eight times its dry volume of water and to swell to a greater extent. It is sticky and plastic when first wetted, but with further additions of water it becomes a soapy paste. Both bentonite and metabentonite tend to be well stratified. Fossils are generally not common, but they may be abundant.

Pure bentonite is largely composed of the clay mineral montmorillonite (Ross and Shannon). Other substances are particles of plagioclase, orthoclase, biotite, quartz, apatite, zircon, magnetite, glass, and agate. Most particles are angular. Impure bentonite contains other sediments of various compositions and origins.

Metabentonite does not expand when placed in water as does bentonite. It occurs in the older geologic systems and has been found to have extensive distribution in the Ordovician of the upper Mississippi Valley (1), the Appalachian Region (Rosenkrans; Whitcomb and Rosenkrans) and over parts of the Cincinnati Arch. The clay mineral in many occurrences has been identified as montmorillonite, in a few instances as beidellite, and in others as a potash-containing clay mineral. Other substances are feldspar, quartz, zircon, and mica.

Bentonites and metabentonites are generally assumed to have been produced by decomposition of volcanic "ash" (Hewett, Ross, Shannon, Rubey), an acid ash generally having been assumed to have been the original form. However, since both beidellite and montmorillonite are formed in decomposition of many substances, the presence of these two clay minerals does not prove a shower of volcanic "ash." Proof of the volcanic origin of a rock identified as bentonite must rest on the presence of glass relict structures and minerals characteristic of volcanic rocks. If these are not found, it should not be stated that the rock indicates a fall of volcanic "ash."

Bentonite and metabentonite derived through alteration of volcanic "ash" are assumed to have been deposited fairly rapidly, and each bed is assumed to represent a single fall of "ash" or a rapid succession of falls (Kay). Deposits containing clay minerals of beidellite and montmorillonite not derived from the alteration of volcanic "ash" were deposited at rates accordant with the methods and environments of deposition.

CLAY AND SILT OF AEOLIAN DEPOSITION

Terrigenous particles of clay and silt dimension are spread widely by the atmosphere and should be present in greater to less quantities in the deposits of all environments. Certain fine-grained deposits known as loess are generally considered to be entirely of this origin. Loess has been variously interpreted, and several different types of materials have at times been included under the term. The present general practice is to limit the term to an aeolian deposit composed of particles of clay and silt dimension.

The particles of which loess is composed have a wide-range of composition and have been derived from many sources. Most are obviously of inorganic derivation, but it seems likely that few or no fine-grained deposits of aeolian deposition are without particles of organic origin. Dust deposited upon water surfaces settles to the bottom to make deposits related to the environmental condition of final deposition, and even if uncontaminated with other materials, the deposits are not termed loess.

The most common mineral in loess seems to be quartz. Feldspars, micas, and other minerals are present in greater or less abundance. Calcite is often an important constituent, and some loess collected in the Driftless Area of southwestern Wisconsin contained 25 per cent calcium carbonate. Most particles are angular.

Loess is characteristically without stratification, and it has the ability to stand for a long time with vertical walls. This is interpreted as due to angularity of the composing particles and to reinforcement of the deposits by tubes and rods of calcium carbonate related to holes made by plant roots and stems which were lined or filled with calcium carbonate as the plant matter decayed.

Unweathered loess has a light-yellow color and a considerable porosity. Fossils are not common, and those present consist of land organisms. Concretionary structures of a wide variety of shapes may be present.

Various opinions have been expressed respecting the sources of the materials of the loess. The loess of the Mississippi and the Rhine valleys has been considered to have had some connection with Pleistocene glaciers, and it has been suggested that the composing materials were derived from outwash plains and the surfaces of glaciers. Leverett has stated that they were largely derived from the semiarid regions east of the Rocky Mountains and that the materials of the European loess also had a semiarid provenance. The loess of China is generally considered to have had its provenance in the dry regions of western China and central Asia, where it may have been acquired from the surfaces of alluvial fans, floodplains, playas, bared slopes, and many surfaces upon which there were fine materials.

Not all have accepted the aeolian origin of loess. The latest dissenter is R. J. Russell, with his views based on studies of the loess of the lower Mississippi Valley. He considers this loess originally to have been alluvium which has become a colluvial accumulation. The author has studied the loess on which Russell's views were based and does not agree with him. Others not in agreement with Russell are Thwaites and Doeglas.

Atmospheric deposits of fine-grained materials are deposited over all surfaces of the regions of deposition. If the sites of deposition are covered with heavy vegetation, there is complete mingling of the sediments with organic materials, much decomposition and solution, and elimination of the evidence of atmospheric deposition. If deposition takes place under conditions permitting later acquirement of the sediments by water, they become mingled with sediments of other derivation and cannot be recognized. This takes place on lowland areas. If deposition takes place on upland areas with limited vegetation under conditions not permitting much aqueous disturbance or introduction of other materials, the atmospherically deposited sediments are little modified except for reworking by organisms and beating of rain. Reworking by organisms and the beating of rain preclude any possibility of stratification.

So far as known, deposits of loess are of comparatively recent origin, and the greatest known occurrences are in the Pleistocene, in which it has extensive development in the Mississippi and Rhine valleys and over parts of China. Several hundred thousand square miles of the present continental areas are covered with loess. The maximum thickness of the loess in the Mississippi Valley is of the order of magnitude of about 30 meters. The maximum thickness of the loess of China approximates 90 meters (Barbour).

No loess has been proved to exist in the geologic column in formations older than the Pleistocene. Suggestions have been made for a loessial origin of certain geologic units in the Tertiary (Matthew) of the western plains of the United States. Lomas suggested that some of the divisions of the Keuper of England (Triassic) might represent fossil loess.

It is extremely likely that deposits of atmospheric dust were made during every period of geologic time, but equally likely that few have been preserved. There is not much chance of long preservation over lowland areas, as the dusts become mingled with other sediments and lose identity. Dusts deposited over upland areas are exposed to the erosional hazards connected with such positions. There is not much chance for burial beneath other deposits, and burial is essential for preservation. On a land in process of submergence the deposits are subjected to subaerial erosion until attacked by the sea. The marine attack may be expected to result in complete destruction of every trace of the deposit. The limited possibility of preservation probably accounts for the absence or at least rarity of loess in the geologic column.

Under what seem to be rare conditions, fine-grained sediments may become aggregated into particles of sand, gravel, and pebble dimension, following which the particles behave as sands and may be built into dunes. The particles may be produced by mud cracking over floodplains, playas, dried-up lagoons, and mud flats. Dune deposits of this kind have been described by Coffey from a locality near the mouth of the Rio Grande. The dune ridges are stated to be several miles long, 60 to 90 meters wide, and 6 to 9 meters high. It does not seem likely that dunes of this kind have much chance of being made a part of the geologic column.

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THE "RED BEDS"

The "Red Beds" form an extensive group of sedimentary deposits over several parts of the Rocky Mountains and the adjacent Great Plains regions of the United States. Age ranges from Pennsylvanian to Triassic. Similar strata in more or less limited development are present in other divisions of the geologic column in different parts of the world, particularly in the Permian and Triassic of Europe.

The typical "Red Beds" consist of claystones, siltstones, and sandstones in which there are thin local units of granule and pebble conglomerate. Interstratified with these clastics are beds of calcium sulphate and limestone. The calcium sulphate strata frequently have extensive distribution, and the range in thickness is from a few millimeters to about 12 meters. The limestones seem to be mostly dolomitic. All beds tend to have considerable variations in thickness, particularly those of the clastics.

The dominant mineral of the "Red Beds" seems to be quartz. Many particles are well rounded and have mat surfaces. Wind has evidently been concerned in the production of these particles. Most particles are coated with a thin film of iron oxide or clay colored with iron oxide. Ferric oxide, or ferric oxide mixed with clay and silt, fills the interstices between the particles of quartz. The characteristic color is due to this oxide, which really forms a very minor part of the sediments. The color in most cases is considered primary or inherited from previous red beds. Krynine states that the brownish-red color of the Supai formation, a red-bed formation of the Colorado Canyon Region, is mostly due to secondary alteration, although some may be inherited from a previous red-bed formation. Fossils may be present or absent in the "Red Beds," the latter far more often the case than the former.

In some cases the "Red Beds" pass laterally into strata of other colors, as in the Permian of Oklahoma, Kansas, and Texas and in the Sespe (Reed) (Tertiary) formation of California. With the change of color there may appear a greater abundance of fossils.

Considerable uncertainty exists respecting the origin of the "Red Beds." As with most sediments, they originally were referred to a marine origin, but the weight of opinion at the present time refers them to deposition in a continental environment which at times had some connection with the sea. There have also been questions raised with respect to provenance of the sediments.

The provenance of the sediments composing the "Red Beds" that were red at the time of deposition depends upon whether the colors are due to original red minerals or to red decompositional products, and in the latter case whether sources of the sediments were preexisting red beds or red residual soils. There are red beds in the geologic column, such as some of the red sandstones and siltstones of the Triassic Newark Series of the Connecticut Valley, of which the colors are partly, and in some cases entirely, due to red feldspar. The typical "Red Beds" owe their colors not to original minerals of parent rocks but to ferric oxide formed as a decompositional product. The parent materials were either older red beds or red residual soils. Older red beds in some cases may have been the parent materials, but red residual soils are considered the provenance of most of the "Red Beds," of which the sediments were red at the time of deposition.

Red residual soils in any considerable extent are formed only on upland areas with warm and moderately moist climate and good underground drainage. In the United States these soils form in Virginia, Tennessee, Alabama, Mississippi, Louisiana, southern Kentucky, and Georgia. Limited vegetation is favorable for their development, but the conditions should not approach semiaridity. The quartz particles of red residual soils are commonly covered with thin films of ferric oxide or clay colored with this oxide, as in red beds. Red soils may be transported by aqueous or aeolian agencies, but the final agent of of deposition for most of the typical "Red Beds" was water, and water is considered responsible for most of the transportation. It is thought that long distances did not obtain between the sources of the red sediments and the sites of deposition, as long transportation would have removed the film of oxide covering the quartz grains and would have incorporated sufficient organic matter to reduce the oxide of iron with elimination of the red color. It is thus concluded, following Tomlinson, that "development of ferruginous soils is the first prerequisite to the development of red beds of the western type," and with Reed that they developed under physical conditions and climates not unlike that of the southern Appalachian Mountains of the present time.

Views respecting the environment of deposition of the "Red Beds" have ranged from marine to continental. It is not unlikely that some parts of the "Red Beds" were deposited in each of several environments. Advocates of a marine environment have pointed to the occurrence of excellent sorting, ripple marks, thin and persistent bedding, presence of marine fossils, lateral gradation into marine strata, and occurrence of beds of calcium sulphate and limestone. Good bedding, ripple marks, sorting, etc., are not proof of marine origin, as these features develop in every environment in which deposition is done by water. Marine fossils indicate probable marine origin for those strata containing the fossils, but not necessarily for any of the associated strata.

Lateral gradation into marine sediments proves nothing except the gradation. Calcium sulphate beds are deposits of evaporating waters containing this salt in solution, but normal marine waters are precluded. Limestones are formed in many environments. Persistence of color in red sediments does not seem probable in a normal marine environment, as organic matter would likely be present to reduce the iron to the ferrous condition. The persistence of the red color indicates that deposition must have taken place under conditions that permitted incorporation of very little organic matter.

Many parts of the "Red Beds," as may be seen in Kansas, Oklahoma, and Texas, are not well sorted or well bedded and show depositional features best interpreted as due to the stream environment, although Krynine (1949, page 65, Fig. 7) states the Permian "Red Beds" of Kansas are marine.

That some parts of the "Red Beds" were deposited in marine waters is accepted as an explanation for those parts containing the marine fossils. However, assignment of particular parts of the "Red Beds" to deposition in a marine environment is no reason for assigning any other part to deposition in the same environment. The calcium sulphate strata were deposited in highly saline waters, and, as the salt is mostly anhydrite, it is assumed that the waters were warm and had a high content of sodium chloride, as indicated by absence of products due to sulphate-reducing bacteria. The waters may have been more or less isolated arms of the sea, or inland lakes to which margining lands did not supply sufficient fresh water to cause outflow. These lands may have been arid to semiarid. The fewness of marine fossils in those parts of the "Red Beds" containing calcium sulphate and limestones accords with the view of high salinity.

Fresh-water lakes as sites of deposition are precluded, as the sediments of these contain too much organic matter, and the sediments could not remain red, if introduced with that color; nor could they become so.

The absence of evaporite deposits and also of marine fossils, as is notably the case for many parts of the "Red Beds," coupled with the presence of sedimentary features seemingly best interpreted as of stream formation, hardly permits the postulate of a marine environment for these parts and suggests deposition over floodplains or the subaerial plains of deltas and possibly to some extent over the subaqueous plains of deltas. Under humid conditions sufficient organic matter would certainly have been incorporated in the sediments in these environments, and loss of red color would have followed. It is possible that deposition may have been so rapid as to prevent much growth of plants, but this rapidity could hardly have obtained over extensive areas. It is hence postulated that deposition took place under climatic conditions not favoring much growth of vegetation. Limitation of vegetation could have been brought about by very low temperature or moderate to high temperature and little rainfall. The former is not possible, as red residual soils do not develop under cold conditions. An extremely arid climatic condition is not a favorable one, as such would permit considerable rolling around of the larger particles by the wind with removal of the ferric oxide films and formation of dunes. Semiarid conditions may permit retention of the red color after deposition unless there is much upward movement of water and the formation of caliche in the surface materials. This might lead to loss of the red color. The most favorable condition for deposition and preservation of color of red sediments on land areas lies between the extremes of some degree of moderate rainfall and semiaridity under relatively high temperature conditions. Deposition need not have been extremely rapid, but rapid deposition would have been a favorable factor. Some parts of the Permian "Red Beds" of Oklahoma, Kansas, and Texas bear evidence of rapid deposition.

The facts lead to the conclusion that many parts of the "Red Beds" were deposited in a subaerial environment with warm and relatively dry climate and that fluvial processes played a large part in the deposition and that other parts were deposited in standing bodies of shallow water of a salinity sufficiently high to prevent activity of sulphate-reducing bacteria. The sites of subaerial deposition were dryer than the regions in which the sediments developed as red residual soils.

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CHAPTER IX

SEDIMENTS OF CHEMICAL DEPOSITION

THE CARBONATE SEDIMENTS

INTRODUCTION

There are many carbonates, but only those of calcium and magnesium are here considered. These two carbonates constitute the dominant constituents of the limestones, so named because lime is made from them by "burning" the rocks to drive off the carbon dioxide and leave lime. Their percentage in sediments ranges from zero to approximately 100. It is proposed to designate as limestones those sedimentary rocks in which the percentage of carbonates is 50 or more. If the percentage is less than 50 the rocks should be given other names with the appropriate qualifying adjectives.

Aragonite and calcite are the two common forms of calcium carbonate. Aragonite is not particularly stable and is thus relatively rare in limestones, although it is a common constituent of shells. Magnesite is the carbonate of magnesium. It does not seem to be an important constituent of sediments, but there are some relatively large deposits. Far more important than magnesite is dolomite, the double carbonate of calcium and magnesium. The ratio of calcium carbonate to magnesium carbonate in dolomite is 54.35 to 45.65, but the two carbonates are not always in this ratio, and in many cases the place of some of the magnesium is taken by manganese, ferrous iron and rarely by a few other elements without any change in crystallographic structure. Excess calcium may be present in the dolomite crystal to a quantity as great as 20 per cent without modification of crystallographic structure; the calcite crystal can contain only about 1 per cent of magnesian carbonate. The calcite crystal can also contain some manganese. Dolomite and calcite are differentiable through their optical properties or more simply by means of treatment with cold hydrochloric acid, calcite effervescing freely and dolomite slowly. They may also be differentiated by means of treatment with Lemberg's solution, calcite turning blue and dolomite remaining unchanged.

There are two chief varieties of limestone of which one is composed dominantly of calcite and termed calcitic limestone, and the other dominantly of dolomite and termed dolomitic limestone. Shrock has proposed to apply the name of *dolostone* to limestones composed dominantly of the mineral dolomite. To be consistent, limestone composed of calcite should then be termed calcistone. As these rocks are now and have generally been known as limestones and probably will continue to be so known and as lime has been made from both of them (whence the name of limestones), the author is of the opinion that it is best to continue such usage with the qualifying adjectives of calcitic and dolomitic. The two varieties of limestone exist in all gradations from those in which the carbonate is entirely calcite to those in which it is entirely dolomite. According to Steidtmann, analyses indicate that limestones do not appear to be common in which the dolomite-calcite ratio lies between 80 to 20 and 20 to 80, most limestones seeming to be either dominantly calcitic or dominantly dolomitic. Pure dolomitic limestone seems to be uncommon. Nearly pure calcitic limestones seem to be common, but a little magnesium carbonate is present in many.

Much information exists respecting the origin of calcitic limestones, but not a great deal is known of the origin of dolomitic limestones. The two varieties are considered separately.

CALCITIC LIMESTONES

Introduction

The most common salt in solution in stream and spring waters is generally calcium bicarbonate. Some calcium carbonate is also present, but the quantity is small as the salt is relatively insoluble, sea water free from carbon dioxide at 16°C. dissolving only 0.0131 gram per liter. Calcium sulphate may also be present. The calcium carbonate in stream and spring waters in the bicarbonate form is to the calcium sulphate in the ratio of about 10 to 1, whereas in sea waters the ratio for the same salts is about 1 to 10. The quantity of carbonates in solution is dependent upon the quantity of carbon dioxide available for placing them in the bicarbonate form. The quantity of carbon dioxide that may be held by water is dependent upon the partial pressure of the carbon dioxide in the atmosphere and the hydrostatic pressure and temperature of the water. Deep waters can retain more carbon dioxide than those at the surface; cold waters can retain more than warm. Thus, bottom waters of the sea, because of low temperatures, can contain more carbon dioxide than the warmer surface waters, and polar waters can contain more carbonates than the warmer surface waters of tropical and subtropical regions. Wells has shown that with the partial pressure of carbon dioxide in the atmosphere at 0.000318, water at 1°C. and 28°C. can contain, respectively, 0.101 and 0.078 gram carbon dioxide per liter. In the Gulf of Mexico the average content of carbon dioxide in the surface water was found to be 0.092 gram per liter with a small increase with depth and a range from 0.088 to 0.100 gram per liter. Water under a pressure of 1 atmosphere and with a temperature of 14°C. dissolves a volume of carbon dioxide equal to its volume, about twice as much under 2 atmospheres, about three times as much under 3 atmospheres, etc.

If there is any change in the partial pressure of the carbon dioxide in the atmosphere, there is a corresponding change in the capacity of the water to contain carbon dioxide and, correspondingly, to dissolve and carry in solution calcium and other carbonates. That there have been changes in the partial pressure of carbon dioxide can hardly be questioned, and the great quantity of carbon dioxide and carbon contained in the sedimentary rocks, equivalent in carbon dioxide to about twenty-five times the mass of the present atmosphere, almost certainly proves variation. The view has been advanced that the early Pre-Cambrian atmospheres may have been largely composed of carbon dioxide (Barrell, MacGregor). The evidence is said to be found in the relative absence of limestones in the very early sedimentary rocks, the large quantity of iron carbonate in early rocks, and, according to MacGregor, in a high content of ferrous iron and low content of ferric iron in the early sediments. With an atmosphere high in carbon dioxide, carbonates could have readily been acquired, transported, and held in solution. Hence limestones could have had scant representation among the rocks until such time as plants became sufficiently abundant to begin to lock up carbon. Deposition of carbonates would have been rapid following such plant development, owing to the immense quantities of bicarbonates in solution.

Surface sea water of the warmer parts of the ocean seems to be saturated with calcium carbonate, a view expressed by Murray a long time ago and proved by the more recent studies of Vaughan Wells, Revelle, Wattenberg, and others. McClendon and Wattenberg have stated that in places it is supersaturated with calcium carbonate. Slight changes would lead to precipitation of calcium carbonate from the supersaturated sea water, and loss of carbon dioxide in either saturated or supersaturated water would produce some precipitation.

Trask has presented data suggesting that the quantity of calcium carbonate in marine sediments increases as the salinity of the surface water rises, and the rate of increase is greatest between salinities of 34 to 36 parts per thousand. Where the salinity of the surface waters is less than 34 parts per thousand, the sediments generally contain less than 5 per cent calcium carbonate; where the salinity is more than 36 parts per thousand, the sediments contain, as a rule, more than 50 per cent calcium carbonate. The precipitation may be due to the increased salinity, and, perhaps, it may be that the increased salinity influences in some way the organic precipitation of calcium carbonate. The studies of Eardley and Pia indicate that the solubility of calcium carbonate rises with increased salinity to a certain limit with sodium chloride and then begins to fall at a concentration of 98.4 grams per liter, but with sodium sulphate it continues to rise to the limits of the experimental work at a concentration of 255.9 grams per liter. The results in sea water were not studied.

Colors of relatively pure calcitic limestone range from white to gray. As most contain various impurities it results that every color may be found in limestones. Many that are light colored on fresh fracture become tan or buff on exposure, owing to oxidation of contained siderite, pyrite, marcasite, or iron silicates.

Textures range from microscopically fine-grained to coarsely crystalline. Some limestones are composed of fragmental particles, and in others none of these seems to be present. Some are largely composed of entire or little-broken shells, and others seem to contain little or no material of organic origin, although the microscope frequently shows small fragments of shells. Some limestones are composed of particles that evidently were sands at the times of deposition. Many limestones are composed of oölites, and there are others that seem to be oölitic, but the particles resembling oölites are small pellets of comminuted matter that may represent excrements.

The materials of which calcitic limestones are composed were transported to the sites of deposition in solution, as colloidal particles in suspension, as particles of silt and sand in suspension, and as sand-size and larger particles by traction.

Calcitic limestones are present in every system of the geologic column since the Pre-Cambrian. They occur as formational units and as units of smaller dimensions interbedded with all other kinds of sediments. Most, if not all, calcitic limestone units grade laterally into other kinds of sediments. Most limestones prior to the Cambrian seem to be dolomitic, and such is also the case for many of those of the early Paleozoic.

Clarke gives the percentage of calcium oxide in the earth's crust as 5.10. Vaughan has shown from compilation of data existing at the time of publication that the percentage of calcium oxide in the sediments of the deep sea is 18.03, or 3.53 times the percentage in the crust. According to Kuenen the percentage of calcium oxide in deep-sea sediments is 15 per cent. This apparrent concentration of calcium oxide in deep-sea sediments must be balanced by a deficiency in some other deposits. Perhaps this deficiency may be in continental deposits and those of deltas. It is also possible that the Pre-Cambrian and Paleozoic sediments of the deep sea were low in calcium oxide. Planktonic organisms, particularly pelagic foraminifera, to which deep-sea sediments owe much of their calcium oxide, did not have much development in the Paleozoic and older seas and did not become abundant until late Mesozoic, and thus there may have been little deposition of calcium carbonate in the deep seas of those times. Kuenen has suggested that prior to the development of abundant pelagic foraminifera and their astonishing activity in utilization of calcium carbonate for making shells, oceanic waters might have been oversaturated with calcium carbonate, and much of the salt might have been directly precipitated without intervention of organisms. Deposits would then have been made mostly over shallow bottoms rather than in the deep areas of the oceans. Deposits would have made fine-grained limestones containing few shells. He states there are many limestones of this kind in the Paleozoic and early Mesozoic parts of the column. Present oceanic waters are generally so low in calcium carbonate that little limestone is deposited directly, and deposits like those that made the great Paleozoic limestones are very rare.

It does not follow that all calcium carbonate precipitated results in the formation of calcitic limestone, as conditions on the bottoms may be such as to lead to solution. Many parts of the bottom of the sea that seem favorable for deposition of calcium carbonate are without deposits of this salt, and Correns has suggested that carbonates precipitated in overlying waters are again taken into solution by currents that are high in carbon dioxide and low in calcium carbonate.

Calcium carbonate deposition is related to depth of water for two reasons. Benthonic organisms thrive most abundantly in the zone of light, as light makes possible growth of plants and hence food for animals. Deeper waters acquire much of the food supply by importation from shallow or surface waters; hence life is less abundant and benthonic shell accumulations more restricted than in shallow or surface waters. Planktonic organisms dwelling in the upper waters sink more or less slowly to the bottom following death. Carbonate shells of these organisms are of delicate structure with much surface in relation to volume and are hence readily dissolved. With progress in depth, waters become colder, the carbon dioxide content higher, and the pressure greater, and hence solution becomes more rapid. Murray and Renard's figures show this relation of decrease of calcium carbonate with depth, and Correns has shown that not only does this relation hold, but also calcium carbonate deposition on the sea bottom is in some degree inversely proportional to the presence of bottom currents.

Origin of Calcitic Limestone

Calcitic sediments may be deposited in seas, lakes, rivers, and on the lands. The greatest extent of deposits is on the bottom of the sea, mostly at depths of less than a few hundred meters, but deposits are made to at least 5,000 meters. Deep water is not essential, and deposition may take place at sea level. Temperature, circulation, and the entrance of argillaceous, arenaceous, and ferruginous materials from the lands are factors of importance. Absence of circulation may lead to poisoning of the waters with hydrogen sulphide and sulphuric acid and elimination of those organisms upon which deposition of calcium carbonate so greatly depends. Entrance of a large volume of terrigenous sediments may likewise eliminate the organisms and effectively mask any calcium carbonate that is deposited. Rapid circulation of water with a high content of carbon dioxide may dissolve any calcium carbonate that is precipitated.

Calcitic sediments owe their origin to physical and chemical processes, and the chemical processes are both organic and inorganic. Physically deposited calcitic sediments may result from calcium carbonate carried in colloidal form with precipitation by means of electrolytes; from calcium carbonate of clay, silt, and sand dimensions carried in suspension in either water or atmosphere; and from calcium carbonate of sand and larger dimension transported by traction in either water or atmosphere.

Deposits of calcitic sediments may consist of materials of sand-and-gravel dimensions, aggregations of shell materials mixed with finely divided calcium carbonate, finely divided calcium carbonate, and any of these calcareous sediments mixed with other sediments of varied origins and characteristics. Many unconsolidated calcium carbonate sediments are known as marl, a term with a considerable range of meaning (Twenhofel). Many lake marls of Wisconsin are devoid of any shell matter and consist of very finely divided calcium carbonate.

The different methods of origin of the calcitic sediments have been arranged as follows:

I. Chemical origin.
  A. Organic.
    1. Accumulation of skeletal and protective structures of organisms.
    2. Vital activities of organisms.
      a. Photosynthesis of plants.
      b. Bacterial processes.
  B. Inorganic.
    1. Mingling of solutions.
    2. Escape of carbon dioxide from water.
    3. Evaporation.
II. Mechanical origin.

These different methods of origin are considered in such detail as seems necessary.

ACCUMULATION OF SKELETAL AND PROTECTIVE STRUCTURES OF ORGANISMS

It seems probable that most calcitic limestones are of organic origin and that most were formed in the sea. The shell and skeletal matter may range in dimension from microscopic to shells as large as a meter in diameter. The shells may be entire or in fragments.

Shells composed of calcium carbonate ordinarily contain some magnesium carbonate, a little silica, a little calcium phosphate, and in some cases a little calcium sulphate. Table 21 shows complete analyses of the shells and skeletal structures of the important groups of invertebrate organisms. The organisms evidently acquire the materials for the shells from substances in solution and possibly also in suspension in the waters in which they live and from which they obtain their food. Experimental work by Murray and Irvine has shown that the calcium may be obtained from any calcium salt in solution.

The data given in the table show that calcitic limestones formed from madreporarian and hydroid corals, annelids, most bryozoans, most brachiopods, and the mollusks have shells dominantly composed of calcium carbonate. Calcitic limestones

TABLE 21. THE INORGANIC CONSTITUENTS OF MARINE INVERTEBRATES

Organisms	Percentages					
	$CaCO_3$	$MgCO_3$	SiO_2	$Ca_3P_2O_8$	$(AlFe)_2O_3$	$CaSO_4$
Foraminifera	77 to 90	1.8 to 11.0	*T* to 15	*T*	*T* to 5.0	0
Calcitic sponges	71 to 85	4.6 to 14.0	*T* to 8	? to 10.0	*T* to 5.7	0
Corals	73 to 99.9	0.1 to 16.0	*T* to 2	*T* to 9.0	*T* to 1.0	*T* to 5.4
Annelids	90 to 92	0.0 to 10.0	0	*T* to 1.0	0	*T* to 0.13
Echinoderms	78 to 93	5.0 to 15.0	0.0 to 10.0	*T* to 1.9	0.1 to 5.2	*T* to 4.2
Bryozoa	63 to 97	0.6 to 11.0	0.2 to 6.7	*T* to 2.7	0.0 to 2.3	0.3 to 2.8
Calcareous brachiopods	89 to 99	0.5 to 8.6	0.06 to 0.5	*T* to 6.6	0.04 to 0.5	0.4 to 1.7
Phosphatic brachiopods	1.2 to 8	0.8 to 0.67	0.5 to 0.9	75 to 92.0	0.3 to 1.2	2.9 to 8.6
Mollusks	94 to 99.9	0.0 to 6.0	0.0 to 2.2	0.0 to 0.9	0.4 to 1.9	0.0 to 0.2
Barnacles	96 to 98	0.8 to 2.5	0.33 to 2.1	0.0 to 0.8	0.15 to 0.7	
Crustaceans	29 to 83	3.7 to 16.0	0.0 to 3.8	8.7 to 27.0	0.05 to 8.9	*T* to 5.3
Calcareous algae	74 to 99	10.9 to 25.0	0.02 to 2.1	*T* to 0.4	0.01 to 1.6	0.03 to 1.4
Average	83.30	6.36	2.02	5.60	14.14	1.11

T equals trace.

Modified and condensed after F. W. Clarke and W. C. Wheeler, *U.S. Geol. Surv., Prof. Pap.* 124 (1922).

formed from the skeletal structures of foraminifera, sponges, alcyonarian corals, and echinoderms may contain considerable magnesium carbonate, and those composed largely of algal secretions may contain a rather high content of magnesium carbonate. The studies of Clarke show that the quantity of magnesium carbonate in shells is related in some way to warm temperature, and that no magnesium carbonate is found in aragonite. Certain brachipods, most of the crustaceans, and the young of certain mollusks have shells containing some calcium phosphate, and these shells may produce calcitic limestones with a considerable content of calcium phosphate.

Thorp's studies of the contributions of the different groups of organisms to the calcareous deposits about the coasts of Florida and the Bahamas are given in Table 22. The arrangement of the table is modified from that of Thorp. It shows that the calcareous algae take first place, and that other constituents are more or less in the same degree of abundance in each locality.

TABLE 22. PERCENTAGE DISTRIBUTION OF DIFFERENT CONSTITUENTS IN THE CALCAREOUS SEDIMENTS

Type of sediments	Florida, per cent average of 50 samples	Bahamas, per cent average of 24 samples
Calcareous algae	25.1	18.0
Mollusks	17.5	12.2
Silt	13.9	11.7
Coral	9.3	8.2
Foraminifera	9.0	17.3
Clay	7.8	14.8
$CaCO_3$	5.3	6.0
Spicules, total	4.3	2.1
Minerals	3.9	0.5
Crustacea	1.4	0.7
Worm tubes	1.4	3.0
Bryozoa	0.4	*T*
Oölites	0.4	1.6
Pellets	*T*	3.8
Aggregates	....	0.8
Total	99.7	100.7

T equals trace.

The skeletal structures of most invertebrates are largely composed of calcite. Some have a part in the form of aragonite, and aragonite composes the inner pearly layer present in many shells. Böggild states that most salt-water shells consist of aragonite and calcite, whereas fresh-water shells are mostly aragonite. The presence of aragonite has some bearing on preservation, as the instability of this mineral leads to its early disappearance or change to calcite.

CONTRIBUTION OF DIFFERENT GROUPS OF ORGANISMS TO FORMATION OF CALCITIC LIMESTONES

Foraminifera. Foraminifera are now very important in the formation of limestone. Some are benthonic and others planktonic. Benthonic forms tend to have relatively large shells, planktonic forms are small. Planktonic forms dwell in all marine surface waters, and their shells may accumulate on all bottoms not so deep that shells go into solution before reaching bottom. Accumulation of the shells on the sea bottom forms ooze of which the most important is composed of the shells of *Globigerina.* Because of solution of the delicate shells in the long passage from the surface to the deep bottom, it results that little remains of shells below depths of about 5,000 meters. Most Globigerina ooze is on bottoms of less depth. This ooze has been estimated to cover 22,500,000 square miles of the bottom of the Atlantic, 14,800,000 square miles of the Pacific, and 12,200,000 square miles of the Indian Ocean for a total of over 50,000,000 square miles. The ooze is rarely composed entirely of shells of *Globigerina.* Present are shells of other planktonic organisms and shells of some benthonic organisms. There is also always some inorganic matter.

The color of nearly pure Globigerina ooze is white to bluish white. Admixture of inorganic constituents leads to a gray color. Pure ooze is very fine grained. The appearance is generally homogeneous except when large benthonic shells are present.

Foraminifera have been known since the early Paleozoic. Benthonic forms first became abundant in the late Paleozoic, and planktonic forms in the late Mesozoic. Since the appearance of planktonic forms there has been a steady entombment of calcium carbonate in the deposits of the deep sea. This represents a permanent loss to the continents and a lessened precipitation in shallow water. Kuenen (1941) states that 10 of the 15 per cent of the calcium carbonate in deep-sea deposits represents the loss from continental areas, and continuance of such loss will lead to a lime famine in about 150 million years.

Shells of benthonic foraminifera are present in considerable abundance in the geologic column beginning in the Mississippian. None of these formed oozes. Beginning with the Cretaceous there are deposits that contain foraminiferal shells like those in modern ooze, but typical fossil foraminiferal ooze is not common. This probably is due to the fact that it is largely a deep-water deposit and hence rarely appears in the geologic columns of the land areas.

Chalk has been interpreted as a fossil ooze (Tarr), although in many cases foraminiferal shells are not present in abundance. Much chalk is composed of very finely divided calcium carbonate, to the formation of which it seems probable that foraminifera made only limited contributions. The environment of deposition was largely shallow water, as shown by the presence of large shells, characters of overlying and underlying strata, and, in some cases, interbedded strata. Much chalk of North America does not contain many organisms and corals and brachiopods are conspicuously rare or absent. Conditions evidently were not favorable for these organisms. Sediments that might produce chalk are the calcareous oozes mantling shallow areas in parts of the Bahamas. These have been termed *drewite*, and they are stated to be composed very largely of aragonite. They have been variously interpreted, but no one has related them to foraminifera. It has been suggested that, if oozes composed of particles of aragonite changed to calcite, a dense, fine-grained limestone would be formed, but, if the particles were composed of calcite, lithification might not take place and a chalk would result (Böggild).

Sponges, Corals, and Bryozoa. These three groups are associated, as all have been responsible at one time or another for formation of reef limestones.

Sponges have not contributed greatly to the formation of calcitic limestones. The widely distributed *Archaeocyathinae* reefs in the Lower Cambrian form one of the notable exceptions. Another exception is found in the reefs forming crags of unstratified limestones in the White Jura of the Swabian Alps. These are partly due to calcareous sponges, but the rocks in large part are dolomitic limestones.

Corals have made large contributions to the formation of limestone since the Ordovician. There are many reeflike structures in the Silurian of Anticosti and the Michigan Basin (Cumings and Shrock; Shrock) of northern United States and southern Canada. Perhaps the finest reefs in the Silurian are on the Island of Gotland (Fig. 43) and the adjacent coast of Esthonia. These reefs are beautifully exposed in the coastal cliffs as large structureless masses of limestone. They form elevated areas in the interior of Gotland of which some are used as sites for churches. The "ball stones" in the Wenlock of England are bodies of reef rock. Coral reefs are present in the Lower Devonian of Bohemia, the Onondaga of New York and parts of the Ohio Valley, and the Middle Devonian of Iowa. Reefs are present in the Mississippian

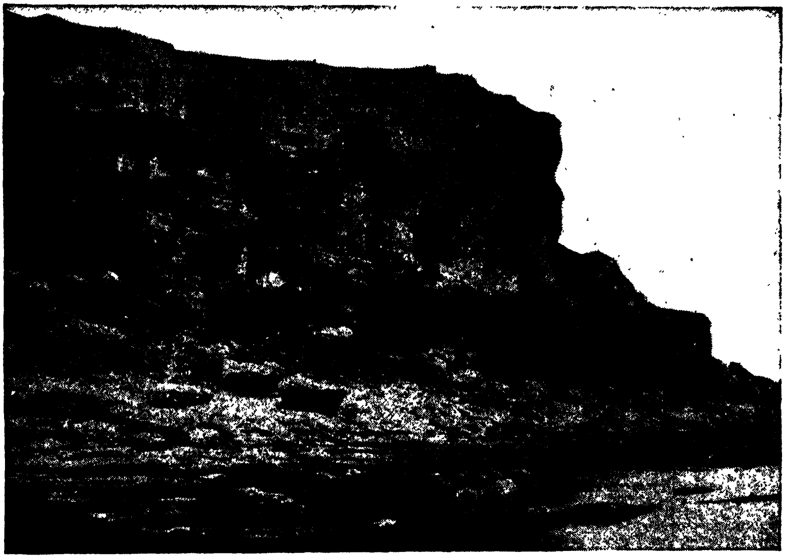

FIG. 43. A Silurian coral reef. The reef forms the promontory near the south end of the Island of Gotland in the Baltic Sea. The rock on the right is the limestone that is formed of the coral reef; that on the left is limestone composed of coral, mud, and sands. The reef rock is unstratified, the other rock has stratification. The irregular contact between the unstratified limestones of the reef and the stratified limestones to the left is shown by the white line. The limestones of the reef consist of coralla of *Favosites*, *Syringopora*, *Halysites*, *Clathrodictyon*, and some others. (*Photograph after Munthe, Swedish Geol. Surv.*)

of Great Britain, and the Jurassic reef limestones of the Solenhofen region of Bavaria are famous the world over because of the excellent preservation of fossils in fine-grained limestones adjacent to the reef rock. At the present time reef limestones have great extent in many tropical waters, and the Great Barrier Reef on the east and north coasts of Australia forms a ridge of limestone on the sea bottom for over 1,000 miles.

Colonial corals have been among the chief contributors to the building of the reefs (bioherms), largely because these have been able to build massive resistant structures. Both individual and colonial forms of corals have aided in forming calcitic limestones in places where reefs do not exist.

Coral reef limestones are without, or have very poor, stratification, and tongues of reef rock extend into the adjacent better stratified sediments. On the outer side are stratified limestones that may be excellently crystallized. Some of these adjacent to one of the reefs of Gotland are known as the Hoburgen marbles.

Coral reefs are places of great organic activity. Between the individual coral colonies are protected spots inhabited by adapted benthos. Corals constitute the food supply for many organisms. The coralline structures by reason of ease of cutting and close proximity to food afford homes for many boring animals, so that reef structures are riddled long before burial removes them from attack. The borers add their skeletal structures and the products of their boring to the reef material. Organisms dwelling on the reef and in spaces between the coral colonies are of many kinds, and in modern waters their contributions to the materials of the reefs surpass those of the corals. According to Howe, algae are the largest contributors to reef materials, and in the boring made at Funafuti it was found that the materials composing the reef in order of their quantitative importance were *Lithothamnium*, *Halimeda*, *Foraminifera*, and corals. Borings made on the Great Barrier Reef show a like relationship (Richards).

The stratigraphic problems connected with coral reefs are complex. The reefs may rise as much as 30 meters above the bordering bottoms and they have nearly flat summits. They determine four more or less distinct environmental conditions: the bottom outward from the barrier, the exposed barrier, the barrier summit, and the protected lagoon. Each of these environments supports an adapted group of organisms, and those on the exposed barrier and the barrier summit may be many meters above those outward from the barrier. The sediments range from the finest of lime muds in the lagoon to coarse gravels on the margin of the exposed barrier and in channels crossing the reef. The rocks of the reef are more or less structureless, those of the lagoon are evenly stratified and more or less horizontal in original position. On the outer side of the barrier the strata may incline seaward at relatively high angles. There may be many local unconformities, and considerable slumping may be indicated. The sediments deposited outward from the barrier are in considerable part derived therefrom and consist of particles ranging in size from that of clay to that of coarse gravel. Many particles of gravel dimension may be more or less worn coral colonies, and particles of sand dimension are composed of fragments of skeletal structures of many kinds of organisms. Sediments connected with coral reefs are estimated to cover about 2,700,000 square miles in the Pacific, 800,000 square miles in the Atlantic, and 400,000 square miles in the Indian Ocean.

It is thought that lateral extensions of coral reefs may indicate halts in rise of sea level. As sea level rises, the reef organisms build upward in an attempt to keep built to the highest level possible. Halts in the rise of sea level more or less compel seaward extensions of the reef, which, with further rise of sea level, become covered with other sediments. The extent of the projections may serve to suggest the duration of the pause in the rise of sea level (Fig. 44).

Coralline skeletons are composed of calcite, but many coral reefs of the geological column and some very recent ones have been altered to dolomite. Obviously this is a case of replacement.

Bryozoa are the so-called moss animals. They have a colonial habitat like many corals, but the secretions of calcium carbonate by an individual are of much less quantitative importance, and, correspondingly, reef structures or bioherms are much smaller. Bryozoan reefs have been described from the Clinton and Niagara of New York, where they occur as relatively small, more or less isolated reeflike masses in the midst of other sediments. These masses are not composed of pure limestone, supposedly owing to the fanlike and dendroid structures of the colonies filtering inorganic sediments from passing waters. Bryozoan remains may be found in nearly all limestones since their appearance in the early Ordovician.

Annelids. Annelids, or wormlike animals, have lived since the Pre-Cambrian, but their contributions to rocks have never been particularly important. Cretaceous strata contain small bodies of limestones made by serpulids, and serpulids build small atoll-like structures on the coast of Bermuda.

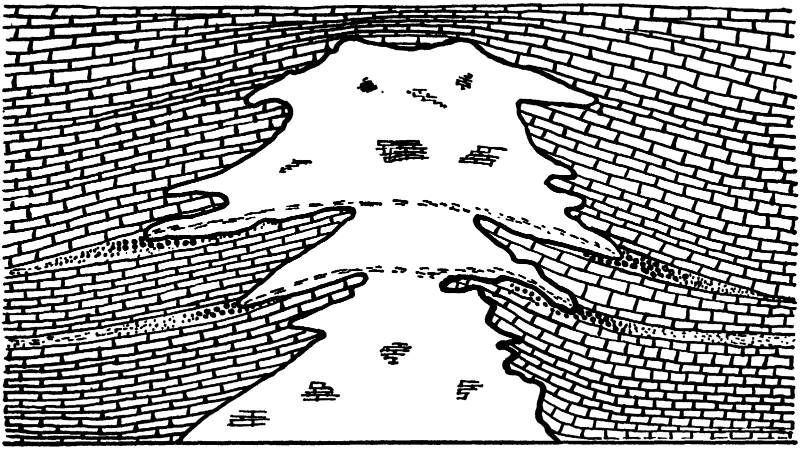

FIG. 44. Diagram showing relations of coral-reef rock to those associated. A stratigraphic break may be assumed to exist just above the places where the reef spreads laterally. This is indicated by the shaly zone through the reef. [*Based on observations of the reefs of Gotland, Anticosti, eastern Wisconsin, and the work of Munthe* (*On the Sequence of Strata within Southern Gotland, Geologiska Föreningens Förhandlingar*, 32 [1910], pp. 1397–1453), *Grabau* (*Principles of Stratigraphy*, 1911, pp. 426–427), *and others*.]

Echinodermata. Of the seven divisions of the Echinodermata, the asteroids, ophiuroids, and holothurioids have made relatively insignificant contributions to the formation of limestone. Cystoids made more or less small contributions in the early Paleozoic, blastoids made little more than minor contributions to limestones during the Mississippian, and echinoids aided considerably in forming limestones during the Cretaceous and Tertiary. Crinoids at times during the Paleozoic made very important contributions to limestone formation. Some limestones are largely composed of crinoidal remains, among which are the crinoidal limestone in the Chicotte formation of the Silurian of Anticosti Island, the *Crinoidal Limestone* of the Silurian of Gotland, and the *Crotallocrinus Kalk* of the Oslo region of Norway. Some of the limestones of the Mississippian and Pennsylvanian systems contain significant quantities of crinoidal remains. For some unknown cause, many crinoidal limestones have a light red to pink color. Crinoids tend to live in colonies more or less to the exclusion of other organisms, hence accounting for the somewhat local occurrences of crinoidal limestones. Skeletons of modern echinodermata contain some magnesium carbonate, and it may be that fossil forms had the same characteristic. Limestones composed very largely of fragments of crinoids have been termed criquina (Tester).

Brachiopods. Brachiopods took a large part in the formation of calcitic limestones from the early Ordovician to the end of the Paleozoic. They have played a minor role since that time. Many Ordovician calcitic limestones are largely composed of orthoids and stophomenenoids, and the *Pentamerus limestones* of the Silurian and early Devonian are largely, and in some cases almost entirely composed of pentameroids. Productoids constitute large parts of some Pennsylvanian and early Permian limestones.

Mollusks. Mollusks have contributed to the formation of calcitic limestones since the Ordovician, but they did not become particularly important until the Pennsylvanian, when certain limestones were largely made from myalinoid shells. The Jurassic and the Cretaceous contain calcitic limestones largely composed of shells of *Gryphaea* and *Exogyra*, and some limestones are composed of oysters. Cephalopods have occasionally made sufficient significant contributions to give names to stratigraphic units, as the *Vaginatinkalk* of Esthonia. Gastropods have had a somewhat similar importance and likewise have given names to stratigraphic units, as the *Gastropodenkalk* of Norway, and the *Worm Rock*, now forming on the coast of Florida, composed of the tubes of a species of gastropod. Mollusks now form large accumulations of shells on the present sea bottom and in places build banks of shells. Broken and poorly cemented shells have generally been termed coquina, and this rock has been rather extensively used on the coast of Florida in construction. Ancient limestones similarly composed of shells have sometimes been termed fossil coquina. Many fossil coquinas, however are composed of the shells of brachiopods.

Most mollusks belong to the benthos, but there are some planktonic forms among which are the pteropods. These have small shells and form ooze over some parts of the sea bottom to a maximum depth of about 3,000 meters. The most extensive deposits seem to be in the Atlantic, where this ooze covers an estimated area of about 400,000 square miles. Pteropods have been in existence since the Cambrian, but their contributions seem to have been minor until the present.

Crustaceans. Crustaceans have contributed to the formation of limestones since the beginning of the Cambrian, but the contributions, although common, have never had more than minor significance.

CALCITIC LIMESTONES CONNECTED WITH VITAL ACTIVITIES OF ORGANISMS

Introduction. Calcitic limestones of this origin are not connected with skeletal structures. The carbonate is precipitated as a consequence of certain vital activities of organisms. The work may be placed in two classes, one connected with photosynthesis of aquatic plants and the other with bacterial processes.

Calcitic Limestones Connected with Photosynthesis of Aquatic Plants. All green plants require carbon dioxide. Aquatic plants satisfy their needs from carbon dioxide in the water and from bicarbonates, reducing these to carbonates, which may then be precipitated. The precipitation takes place upon adjacent surfaces, thus covering leaves, stems, and the bases to which the plants are attached. The precipitated carbonates are in very finely divided form and seem to consist of small crystals of calcite to which Wood applied the name of *algal dust*.

Work of this kind is done to all depths in which plants containing chlorophyll grow. Green plants are known in the Caribbean to the depth of 284 meters, but it is thought that the lower limiting depth of much carbonate precipitation by photosynthesis is about 80 meters. One of the most common algae concerned in the precipitation of carbonates is *Lithothamnium*. This alga is extremely abundant in most oceanic waters, and locally its deposits cover everything with a white coating or make globular balls resembling those of popcorn. *Lithothamnium* is an extremely important contributor to the coral reefs, and in the boring on Funafuti it was found to be the most important contributor. *Halimeda* is another extremely common form that was found to be second in importance in the Funafuti core (Finckh). This alga grows with great rapidity, and one measured on Funafuti attained a height of 55 mm. and a diameter of 80 mm. in 6 weeks and precipitated 1.38 grams of calcium carbonate in that time (Gardiner). *Lithothamnium* has been stated to be the principal builder (Howe) of the Bermuda reefs, with the corals taking second place, and a somewhat similar relationship seems to hold in other regions where studies have been made (Thorp). Lohmann has shown that under each square meter of the North Atlantic there live about 500,000,000 of the small lime-secreting plants known as coccoliths. These divide about every third day so that about one-third die and sink toward the bottom every day. As the individual organism is very minute, the annual accumulation beneath each square meter is not large, but when the entire area of the oceans is considered, the quantity of calcium carbonate precipitated by this small organism becomes of astronomic magnitude.

Green plants, many algae, but also many higher plants do the same kind of work in fresh waters. Studies have been made of this work in many parts of the world and particularly in lakes of Wisconsin under the leadership of Doctors E. A. Birge and the late C. Juday. *Chara* in Green Lake contains about 70 per cent calcium and magnesium carbonates on a sand-free and air-dry basis of which only about 2.5 per cent is magnesium carbonate. As the waters of the lake contain about 1.4 times as much magnesium carbonate as calcium carbonate, it is obvious that the plants are precipitating little magnesium carbonate (Rickett). The growth of *Chara* is said to average about 800 lb. per acre of air-dry plant matter over the shallow parts of the bottom on which it grows, and it is estimated that this plant annually deposits about 1,000 tons of calcium carbonate in Green Lake. This, at the place of maximum deposition, averages about 1 cm. of thickness in 100 years. As *Chara* is stated to be about half the annual crop of green aquatic plants, the rate of deposition due to plants is increased to about 1 cm. in 50 years. Carbonate that did not lodge upon the plants was not measured and, as other agents of precipitation may be present, it is considered that the rate of deposition may be as great as 1 cm. in 30 years.

So far as plants are concerned, the carbon dioxide required for photosynthesis may be acquired from any bicarbonate in solution, and if several are present some of each carbonate may be precipitated, depending upon equilibrium and solubility relations and the ease with which plants may detach carbon dioxide from the bicarbonates. Green Lake in Wisconsin contains more magnesium carbonate in solution than calcium, and Lakes Mendota and Monona have the two carbonates in about the same ratio as in dolomite. If the quantity in solution determined the precipitation, there would be about as much magnesium carbonate precipitated as calcium carbonate. But magnesium carbonate is more soluble than calcium carbonate, so that the latter is deposited, whereas most of the magnesium carbonate remains in solution and is carried from the lakes. The result is that the marls in the lakes contain very little magnesium carbonate. As magnesium salts are much more abundant in marine than in fresh water, whereas calcium carbonate is much less, it is possible that considerable magnesium carbonate may be precipitated by algae in marine waters with the calcium carbonate. However, the work of Wattenberg and Timmermann shows that sea water is not saturated with magnesium carbonate, thus making precipitation difficult.

Limestones originally due to photosynthesis are extremely abundant in the geologic column (Johnson). Reeflike bodies and smaller units referred to this origin are locally present in great abundance in the Huronian of Canada and the United States. Limestones of this origin are also abundant in parts of the Cambrian and Lower Ordovician. Occurrences are not so extensive but are common in the Middle and Upper Ordovician, Silurian, and later Paleozoic systems (Garwood). There is great development of reefs of algal limestone in the Permian of west Texas (King) and southeastern Russia. The Triassic of the Alpine region contains reef limestones of algal origin with more or less continuous accumulation through a thickness of 900 meters. Many of these algal limestones are now dolomite (Skeats, Ogilvie-Gordon). There are also immense accumulations of limestones which are not in the form of reefs.

The reef rock of algal limestones, except for wavy laminations connected with growth, is more or less without bedding. There is often a brecciated appearance. Inclinations about the margins are frequently very high, and there is often a considerable development of oölite and pisolite.

More recent algal deposits are the reeflike bodies in the Tertiary Green River formation of the Rocky Mountain Region and the tufas of ancient Lake Lahontan of Nevada (Russell). The Lahontan tufas are of three varieties: lithoid, thinolitic, and dendroid. They form more or less domelike masses of variable shapes, of which some are towerlike bodies 15 to 18 meters high. The maximum thickness is about 18 meters, with 6 meters an average maximum. The calcium carbonate content is a little less than 90 per cent, and there is between 4 and 5 per cent magnesium carbonate. The domal structures have a central part that is composed of lithoid tufa. This has a covering of thinolitic tufa that in turn is surrounded by dendritic tufa, which constitutes the major part of each dome.

Lithoid tufa is gray and occasionally contains shells of gastropods. It was the first to form and is hence the oldest. It is present in the form of beds and cores of domes. Thinolitic tufa was second in order of formation. This tufa is composed of prismatic structures up to 15 to 20 cm. long and 1 cm. thick. These were at one time considered crystals. Thinolitic tufa ordinarily is 1.8 to 2.4 meters thick but near Pyramid Lake there is a thickness of 3 to 3.5 meters. Dendroid tufa is also gray, and it occurs as spheroidal, mushroom, and dome-shaped masses with concentric macrostructure and less pronounced internal dendritic structure. There are several square miles covered with domal structures, up to 1.5 to 1.8 meters in diameter, that are largely composed of this variety of tufa.

Recent deposits due in part to photosynthesis are found in many hard-water lakes of the Temperate zones. Accumulations to 9 or more meters in thickness are known. The deposits are generally known as marls, of which only a part may be of algal origin (Davis). Wisconsin has extensive occurrences in the southern part of the state. The marls of Green Lake, which is over 70 meters deep and the deepest lake in the state, cover the deepest as well as the shallower bottoms. These marls are composed largely of shells. "Algal biscuits" or "marl balls" are common in some lakes, and they are extremely abundant locally in spherical and disk-shaped specimens up to 15 cm. in diameter.

Some resemble pop-corn balls. Mawson has described "algal biscuits" from occasionally flooded depressions in southwestern Australia. The "biscuits" range in dimension from very tiny to 20 cm., and they are so abundant in places as almost to form an "algal biscuit" pavement.

The Great Salt Lake and Red Sea oölites were referred to algal origin by Rothpletz. Perhaps the explanation is valid, but it does not seem necessary to resort to algae to explain these structures, although it is possible that they may be responsible in some cases.

Many algal limestones are highly dolomitic, and it is possible that they were so in the beginning. This is the case for most Huronian, Cambrian, and Ordovician occurrences and also for the great algal reefs of the Alps.

Calcitic Limestones Due to Work of Bacteria. Drew suggested and attempted to prove that denitrifying bacteria are indirectly responsible for certain fine-grained calcium carbonate sediments that have extensive distribution at Tortugas and were actually observed by McClendon (1918) in process of precipitation. Bacteria of the kind held responsible have been identified in many parts of the sea and also in Great Salt Lake. They do not live in the surface or overlying waters but in the fine-grained sediments of the bottom, in which as many as 160 million per cubic centimeter have been counted in muds collected on the west side of Andros Island in the West Indies. Kellerman and Smith found that ammonifying bacteria may be among the agents responsible for the precipitation of calcium carbonate, but that organic matter must be present to aid in metabolism before there is much precipitation. Smith placed the bacteria found in sea water and sea bottom muds in six main groups, among which are denitrifiers and strong ammonifiers, which dominate in the muds of the bottom where their activities ultimately lead to the formation of calcium carbonate from calcium sulphate. He concluded that the more abundant the food the greater the formation of calcium carbonate, and that in the muds there is sufficient food for the formation of only a small quantity. Bavendamm found relatively few bacteria in the white calcium carbonate muds covering the sea bottom and banks west of Andros Island, but found that in muds of the coast and in mangrove swamps bacteria were abundant, the surface muds of the mangrove swamp on Williams Island containing 16 million bacteria per gram and, at the depth of 4 ft., 2 million per gram, leading to the view that the formation of the carbonate takes place in the muds of the mangrove swamps and similar places. Field expressed the opinion that marine bacteria may not be important in the formation of calcium carbonate, that there is no direct evidence that the carbonate muds off Andros Island are of marine precipitation, and that they seem rather to be of brackish or fresh-water origin. However, Carey and Waksman have demonstrated that nitrifying bacteria are native inhabitants of marine muds and that nitrite-forming bacteria are present in practically all samples from the bottom. ZoBell has assembled available data and states that aerobic and anaerobic bacteria are present in all marine bottom deposits and are most abundant in the topmost few centimeters and that both types decrease in number with depth. Numbers range to several million per gram of wet mud.

Calcium carbonate-depositing bacteria have been identified in mine waters of the Kimberly region of South Africa on the basis of slides of the calcium carbonate deposited showing bacteria, but no decisive evidence was presented proving that bacteria were responsible for the precipitation of the carbonate seen therein. Williams and McCoy isolated aerobic bacteria in the calcareous muds on the bottom of Lake Mendota, a hardwater lake in southern Wisconsin, that precipitated calcium carbonate from artificial mediums and from the muds of the lake bottom, but the studies failed to prove any connections between the microorganisms in the muds and the calcium carbonate of which the muds are in large part composed, although the calcium carbonate crystals precipitated in the laboratory by the bacteria are like those in the muds.

The bacteria studied by Drew were stated to have their optimum temperature at 29.5°C. (76°F.) and to cease activity at 10°C. (50°F.). Drew postulated that the nitrates considered essential for the nitrate-reducing bacteria may have been washed from the land, may have been produced by decomposition of organic matter, or perhaps may have been formed by nitrogen-fixing bacteria. Study of the water by McClendon (1918) of the region of the assumed activity shows only a very small quantity of nitrate. The smallness may have been due to its utilization by nitrate-reducing bacteria almost as soon as it was formed.

Three ways have been postulated by which bacteria may precipitate calcium carbonate. These are as follows:

1. Nitrates in solution are first reduced to nitrites, which ultimately are changed to ammonia. This reacts with carbon dioxide to form ammonium carbonate, which, reacting with calcium sulphate in solution, leads to precipitation of calcium carbonate. It was stated by Linck that this calcium carbonate is in the form of aragonite. This may need confirmation, although the fine-grained carbonate studied by Drew was reported to be in that form. The bacteria studied by Williams and McCoy formed calcite. Magnesium carbonate may be formed in the same way, but the reaction is stated to be much slower and to take place after most of the calcium has been precipitated. That ammonia is formed in decompositional processes is well known, and there is little doubt that some calcium carbonate may be precipitated in this way. The equations of the reactions are:

$$2NH_4OH + CO_2 = (NH_4)_2CO_3 + H_2O$$
$$(NH_4)_2CO_3 + CaSO_4 = CaCO_3 + (NH_4)_2SO_4$$

2. Ammonia may act directly with calcium bicarbonate, combining with the carbon dioxide and precipitating the carbonate as follows:

$$2NH_4OH + H_2Ca(CO_3)_2 = (NH_4)_2CO_3 + 2H_2O + CaCO_3$$

3. Organic salts of calcium are used by bacteria for food. This releases calcium oxide, which may unite with carbon dioxide and precipitate calcium carbonate.

In spite of the fact that both fresh-water and marine bacteria appear to precipitate calcium carbonate, it does not seem to have been demonstrated that they are in any way responsible for the large volumes of fine-grained calcareous sediments that now exist over some parts of the bottoms of existing seas and lakes, or for similar deposits that existed during geologic time. It can not be doubted, however, that environmental conditions favorable for the precipitation must have often obtained, and must obtain over many areas at the present time, just as the proper conditions can be made to exist in laboratories, and one may confidently assume that bacteria are precipitating fine-grained calcareous sediments now and may have done so from the time of the early Pre-Cambrian. It should be noted that the precipitation is an indirect process and is not directly connected with the vital activities of the organisms but is a by-product of those activities.

The calcium carbonate thought to be precipitated by bacteria, so far as known, is in very small particles and hence in favorable form for participation in the various diagenetic reactions that may take place in the sediments of which the particles are a part.

CALCITIC LIMESTONES FORMED BY INORGANIC CHEMICAL PROCESSES

Limestones may be formed by chemical reactions that in no way are related to organic activity, as follows: through mingling of solutions that produce reactions leading to precipitation; through escape of carbon dioxide from water due to rise of temperature, decline of pressure, or agitation of the water; and through evaporation and rise in salinity. Little is known with respect to precipitation of calcium carbonate through mingling of solutions except where fresh waters loaded with calcium carbonate enter bodies of water of high salinity. This is considered under the topic of limestones formed by evaporation.

It is thought that loss of carbon dioxide from water is of great importance in precipitation of finely divided carbonate, and it is known that calcium carbonate is precipitated on evaporation.

Calcitic Limestone Due to Escape of Carbon Dioxide. Granting essential saturation of sea water with calcium carbonate, it is obvious that any change effecting a decrease in the carbon dioxide content of the water would lead to precipitation of calcium carbonate. A decrease could be brought about by agitation, decrease in partial pressure of the atmosphere, rise of temperature, or decrease in hydrostatic pressure. Agitation is always more or less present and, except locally, probably has little effect on deposits in the sea except in very shallow water, as over deep bottoms the precipitated calcium carbonate would be compelled to sink into cold waters with plenty of carbon dioxide and thus return to solution. It has considerable effect in streams. The partial pressure of the atmosphere may vary from 25 to 30 per cent and probably is a factor of little importance except over shallow bottoms. Changes of temperature may be brought about by cold currents flowing into warm regions; by upwelling of cold bottom waters in warm regions, thus not only raising the temperature but also decreasing the hydrostatic pressure; and by local production of heat through extrusions of volcanic materials on the bottoms of the seas and lakes.

Johnston and Williamson state that, if the carbon dioxide content of the atmosphere in contact with the surface of a water body changes from 3.3 to 3 parts per ten thousand, a change that may take place at the present time, the solubility of the water for calcium carbonate is changed from 65 to 63 parts per million, and 2 parts per million are precipitated. Precipitation of an equal quantity takes place with rise of temperature of about 2°C. Polar waters with a surface content of carbon dioxide sink and move slowly equatorward. There probably is acquirement of calcium carbonate to the limit of saturation in the passage. At the equator the waters rise with rise of temperature and decrease of hydrostatic pressure. Calcium carbonate is certain to be thrown out of solution, and if not dissolved on the way to the bottom a deposit is certain. Johnston and Williamson state that, if a current of sea water with a temperature of 15°C. and saturated with calcium carbonate travels 1,000 km. (600 miles) at the rate of 16 miles per day and attains the temperature of 20°C. at the end of the journey, there would be made 10 changes of water each year over the area traveled. In warming the water from 15 to 20°C. there would be precipitated 5.4 parts of calcium carbonate per million by weight and 2 parts per million by volume, which would make a deposit over the bottom for each passage of a current 100 meters thick, of 2/1,000,000 of the depth of the current or 0.2 mm. This is a rate of deposition of 1 ft. in about 150 years. It cannot be assumed that deposition would be uniform over the entire area, as a current would deposit more at one place than another, some might pass into solution again before attaining the bottom, or there might be large production of carbon dioxide locally, thus overcoming the effects of the change in temperature. Kuenen has suggested that before the advent of pelagic foraminifera sea water may have been oversaturated with calcium carbonate. If that were the case, any rise in temperature would have produced precipitation.

There are many places in the sea where cold waters from the deep bottoms are compelled to rise and flow over shallow bottoms, as some of the banks in the East and West Indies. It would seem that calcium carbonate would be precipitated and deposited on the sides of the banks where the currents rise and over the top unless the precipitates were transported in suspension to the leeward side. Up-wellings on any coast should produce the same result. It is considered very probable that some of the finely divided calcium carbonate deposits found on existing sea bottoms of the tropical parts of the sea are of the origin thus described.

Extrusions of lavas or hot liquids and gases can not fail to heat the waters with which they come in contact and to expel carbon dioxide as a consequence. Precipitation of calcium carbonate is likely, and this precipitate would be deposited in association with the submarine flows. The calcium carbonate in association with some of the submarine volcanic flows in the Ordovician strata of northern Newfoundland may be of this origin. Kania has estimated that a lava flow with an area of 20 square miles and a thickness of 3 meters when accompanied by the usual gas emissions, may lead to precipitation of calcium carbonate equal to 191 circular lenses each 150 meters in diameter and 15 meters thick, and, if a given area of the sea bottom was in some way heated by volcanic activity for a million years, bodies of limestone many meters in thickness could be formed covering an area approximately equal to the area over which the water was heated.

Some calcium and other carbonates are precipitated about springs where the issuing waters lose carbon dioxide because of release of pressure and agitation, and also to some extent through activity of algae living in the water. Deposits of this kind are particularly common about hot springs. The rate of deposition may be very rapid, as at San Vignone in Tuscany where deposits accumulate at the rate of 15 cm. annually and at San Fillippo in Sicily at the rate of 30 cm. in 4 months. Porous deposits of this origin are known as *tufa* or *sinter*, and as *travertine*, if compact and banded. Some tufa is deposited about plants and gives rise to "petrified moss."

Places most favorable for precipitation of calcium carbonate because of agitation are adjacent to places of fast water as falls and rapids in streams carrying hard water (Ohle). In some cases the precipitation is aided by evaporation. The deposits are either tufa or travertine. The Tivola River of Italy deposits travertine very rapidly because of agitation. The extensive Catinga limestone of Brazil originated largely in this way but may also have been aided by rise of temperature and some evaporation. The limestone has a thickness ranging to about 30 meters. Deposition is in progress and takes place downstream below places of agitation. The deposits ultimately form barriers, thus checking agitation above but developing it elsewhere. Bluffs of calcium carbonate have been built to heights of 27 meters. These frequently overhang, and from time to time overhanging parts break off to form talus and ultimately limestone breccia below. Stratification in this type of stream deposit ranges from a *low angle to vertical.*

Calcitic Limestones Due to Evaporation. Limestones due to evaporation are evaporites. The carbonates are among the first to be precipitated as saline waters are evaporated, and theoretically a bed of calcium carbonate should lie at the base of every deposit of evaporites.

Related to evaporation is the precipitation of calcium carbonate that takes place where stream waters carrying a high content of calcium carbonate enter bodies of water of high salinity. Tributaries entering Great Salt Lake bring calcium carbonate to the lake, which contains little because of the high salinity. The carbonates are precipitated about the mouths of the entering streams. Similar deposits are reported to be made where the Jordan River flows into the Dead Sea and, as the concentration of this water body is very high, sulphates are also precipitated at the same place.

Stalagmites, stalactites, and similar structures of caves are deposits of calcium carbonate largely due to evaporation. In the stalactites the materials are banded concentrically to the hollow or porous center. The stalagmites are also banded but parallel to the surface at the time of deposition of each band. The deposits have the characteristics of travertine, but some are massive and are termed onyx (Allison).

Tepetate is a deposit of the arid southwest that consists of calcium carbonate laid down over floodplains and alluvial fans. It is in the form of a thin crust upon the surface or a cement in the surface materials. *Caliche* (*reh* of India and *sabath* of Egypt) is a deposit of calcium carbonate and other salts made in semiarid regions. The salts are brought to the surface by rising waters, which on evaporation deposit the salts on the surface or in the surface materials. Caliche is also termed hardpan in some localities. The deposits and associated cemented materials may attain thicknesses of a meter or more. They conform to the surface, and as they are formed over uplands as well as lowlands they give the impression of strata in inclined or arched position (Breazeale and Smith; Price).

CALCITIC LIMESTONES OF MECHANICAL ORIGIN

Sands derived from calcitic limestone terranes mantle the beaches of many coasts and the bottoms of the adjacent waters. Some beaches on Anticosti and Gotland are composed of sands of which fully 90 per cent consists of particles of limestone derived from terranes of adjacent coasts. Sands of some streams are composed of particles of limestone. Many ancient limestones have ripple marks of aqueous origin, and many are cross-laminated. These structures are known to form only in sands, thus indicating mechanical deposition of these strata. Sands composed of limestone particles may also be deposited by winds.

Many shell deposits are of mechanical deposition. Wave action and scavengers reduce shell matter to sand dimension, after which the particles may be deposited in water or on land. Dunes on Bermuda are composed of shell and coral particles of sand dimension. An older consolidated deposit of the same materials is known as sandstone. The *Miliolitic* formation on the shores of the Arabian Sea is composed of the shells of *Miliola* and is interpreted as a dune deposit, and the Junagarh limestone overlying the Deccan trap in the Kathiawar district of India is considered to have had a similar origin (Chapman, Evans). Dune "sands" on the coast of Galway in Ireland are composed of the shells of foraminifera. So-called coral sands, composed of the skeletal structures of a great variety of organisms, margin coral reefs and shores of abundant shell life. Oölites are also transported and deposited as sands by both water and wind.

Considerable very finely divided calcium carbonate of varied provenance is carried in suspension, and considerable material of this character is spread over parts of the sea bottom. Much of the so-called coral mud produced through comminution of skeletal structures of corals and associated organisms was deposited in this way.

Calcium carbonate of sand dimensions has the characteristics of other sand particles and deposits have porosity and permeability related to packing, sorting, cementation, and sizing. The relatively high content of calcium carbonate in some waters may lead to filling of the interstices between the particles with calcium carbonate, with consequent reduction of porosity and permeability. On the other hand there may be *removal of carbonate with increase in porosity and permeability,* and there seem to be no particularly good reasons for not believing that much original porosity and permeability have been retained and even increased. There are limestones in the Chazy of the Mingan Islands which were "sands" at the time of deposition, and they are not even well cemented in many places today. The so-called coral sands lie adjacent to coral reefs, and "sands" of a somewhat similar character must have lain adjacent to ancient coral and algal reefs, such as those in the Cambrian, Ordovician, Silurian, Devonian, Permian, and Triassic, so that considerable original porosity and permeability may be present in limestones adjacent to these ancient reefs. The reef rocks may also have considerable original porosity and permeability, as the organisms periodically seal up all parts abandoned. Permeability may be low and doubtless would be much lower were it not for the boring organisms having provided access to, and connection between, the sealed cavities.

Summary

The foregoing pages show that calcitic limestones are formed in various ways and under various environmental conditions, ranging from the one extreme of dune deposition to deposition on the sea bottom in depths of thousands of meters. It seems likely that there are few limestones which do not contain constituents derived from several sources and precipitated in several different ways. Some limestones contain an abundance of shells cemented by a limestone "paste" of which the origin is often not determinable. Other limestones are fine-grained and unfossiliferous even in thin section, and it is unlikely that any part of these could have been derived from the skeletal and supporting structures of organisms. Perhaps they were formed by photosynthesis, bacterial activity, or escape of carbon dioxide. Each limestone constitutes a distinct problem that may be solved only on the basis of its characteristics and, to some extent, features in the associated strata. It is probable that most limestones are complex in regard to the agents responsible for the precipitation of the calcium carbonate of which they are composed, and rarely should it be possible to refer limestone to deposition in a single way.

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DOLOMITIC LIMESTONES

Introduction

Dolomitic limestones are largely composed of the mineral, dolomite. They frequently contain some calcite, and there are probably all gradations between completely calcitic and completely dolomitic limestones. In many cases some magnesium of the dolomite crystal is replaced by manganese or ferrous iron without modification of the crystallographic structure.

Most dolomitic limestones are light colored on fresh fracture, but become buff or tan on exposure, owing to oxidation of ferrous iron in the dolomite crystal, or siderite, pyrite, or marcasite in the rock. This change of color is not peculiar to dolomitic limestones; it is commonly present in calcitic limestones.

The textures of dolomitic limestones range from very fine grained to very coarse grained with most in the fine to medium part of the range. Many dolomites on fresh fracture are dense and compact. Some are full of small cavities, and many develop small cavities after exposures of a few years. It has been suggested that these small cavities are due to decrease in volume attendant upon dolomitization of calcitic sediments. If dolomitization were due to replacement of a part of the calcium in calcitic sediments after they had become sufficiently solid so as not to collapse, there would have been shrinkage equal to 12.3 per cent. In many cases it can be demonstrated that the small cavities represent solution connected with removal of fossils or particles of calcite, and it may be proved in other cases that they are cavities connected with deposition. Some dolomitic limestones resemble breccias, and in some cases these may have been formed from algal crusts. What appear to be somewhat similar limestones in England have been described by Dixon and Vaughan and also by Wood as "pseudobreccias." Particles of irregular shapes are surrounded by a matrix of lighter color and coarser texture in the rocks described by Wood, who explains the "fragments" as due to recrystallization of an aragonite portion of an original lime mud accompanied by a somewhat concretionary aggregation of the recrystallized materials to make a pseudofragment. The several authors also state that the matrix is more argillaceous than the "fragments" and that the carbonate in the matrix may be wholly or partly calcite or dolomite. Dolomitization is supposed to be secondary.

Dolomitic limestones are present in every geologic system. They have greatest distribution in the Pre-Cambrian and early Paleozoic systems, but some of the most extensive and thickest dolomitic limestones are in Mesozoic parts of the column.

Dolomitic limestones have a somewhat regional distribution. The limestones of the Ordovician and Silurian of the upper Mississippi Valley are dominantly dolomitic; those of the Cincinnati Arch, Gotland, and Anticosti are dominantly calcitic. Many Mesozoic limestones of the Alpine Region are dominantly dolomitic. Some have referred these differences to hydrothermal alteration, but there is little or no supporting evidence in favor of the explanation. The Silurian and Ordovician sections of the regions of calcitic limestones are commonly thicker than those of the equivalent dolomitic limestones, but on the other hand the Alpine sections of dolomitic limestones are extremely thick. Shallow water is indicated in the sections of Ordovician and Silurian strata. It has been suggested that some of the Alpine limestones were deposited in deep water, but it is thought there is little support for this view.

Some dolomitic limestones contain few fossils, whereas others have them in abundance. In many cases the fossils exist only as molds, and in others they are silicified. Not uncommonly they are composed of dolomite, and rarely calcite is the composing material. There are few small fossils, and the dolomitized fossils show little or no internal structure. It was pointed out by Steidtmann (1917) and again by Wallace that fossils of which the original composition was aragonite are less apt to be well preserved in dolomitic limestones than are those originally composed of calcite. Benthonic forms seem to be those most often present. No careful studies have been made to learn the horizonal distribution of the fossils in dolomitic limestones. This should be done. The fossils are like those in equivalent strata of calcitic limestones. This has generally been considered to signify that the environmental conditions existing at the times the fossils lived were maintained throughout the deposition of the dolomitic limestone unit in which the fossils occur. This interpretation is not justified and can be made to apply only to the particular bed or beds in which the fossils are present. The appearance of the organisms found fossilized may have been mere incidents in the depositional history. At other times conditions may have been difficult or prohibitive for benthonic life.

Nodules of chert and flint are common in many dolomitic limestones, but there are as many dolomitic limestones in which chert and flint are not present. Many calcitic limestones are also filled with chert and flint nodules. The question has been raised as to whether chert and flint are more abundant in dolomitic than in calcitic limestones. This question can not be definitely answered, but flint and chert seem to be about equally common in each.

Occurrence of Dolomitic Limestones

Many dolomitic limestones have formational magnitude, as the Oneonta and Shakopee dolomitic limestones in the Ordovician of the upper Mississippi Valley, the Ordovician Romaine and Beekmantown formations of the north end of the Appalachian geosyncline, the Niagaran limestones of the Michigan Basin, and many others. The range in thickness is to several hundred feet. So far as is known, none of the formations named has been subjected to hydrothermal action, and the positions are not greatly different from what they were at the times of deposition. Many of these dolomitic limestones are filled with dome-shaped algal structures; others have none. Some contain coral reefs or bioherms.

Dolomitic limestones are interbedded with sandstones, siltstones, claystones, calcitic limestones, and various evaporites. Units range from a single bed to a formation, and the boundaries are often sharp. The interbedding has little significance with respect to the origin of the dolomitic limestones, except in those cases where the interbedding takes place with calcitic limestones, as is the case in the Platteville and Decorah formations of the upper Mississippi Valley. Similar interbedding of calcitic with dolomitic limestones has been described in the Pre-Cambrian of Montana and British Columbia, in the Niagaran and Devonian of Iowa, in the upper Jurassic of Germany, and elsewhere.

In many cases the relations can not be described as interbedding as the contacts between the calcitic and dolomitic limestones do not follow, but cut across, bedding planes. It is obvious that in these cases the dolomitization is secondary and not a consequence of original differences in sediments.

Many coral reefs are more or less composed of dolomitic limestones and some are of very recent date of formation. Reefs that are partly of coralline origin are present in the Niagaran limestones of the Michigan Basin, in the great Schlern dolomitic limestones (Triassic) of southern Tyrol, and elsewhere. The coralline, algal and some other organic constituents of these reefs originally may have carried little or no magnesium, and they now have a large content of that element. This is evidently a case of secondary formation of dolomite. In some cases dolomitized reefs pass laterally into calcitic limestones. Many reefs contain few fossils. This may have been caused by destruction produced by diagenesis, or the reefs may have been constructed by algae.

There are many calcitic limestones that contain small to large patches or nests of dolomite. The patches have a slightly different color from the parts composed of calcitic limestone, and the general appearance has commonly been described as mottled. On weathered surfaces the patches of dolomitic limestone are usually yellow to buff in color; the calcitic parts are gray. The dolomitic parts on fresh surfaces are dark, and weathered surfaces usually have a sandy appearance.

Dolomitization in those rocks containing dolomitized patches seems to have spread outward from centers composed of organic and perhaps inorganic materials. Birse has shown that the dark dolomitic spots of mottled dolomitic limestones in the Ordovician of Alberta have central areas resembling tubes, which seem best interpreted as borings of wormlike organisms. Wallace had originally suggested seaweeds. Dolomitization seems to have extended outward from the tubes. Certain fossils in these limestones are composed of calcite, others of dolomite. They originally were largely composed of calcium carbonate.

Large dolomitic masses in the form of lenses and chimneylike and other bodies have been found in limestones in many parts of the world, among which are the Ladinac (Triassic) limestones of the Alps, the Traverse (Devonian) formation of Michigan, and the Mississippian limestones of Illinois and Iowa.

Summarizing the data with respect to occurrences of dolomitic limestones, it is possible to place them in two classes: (*a*) formational and smaller units regularly interbedded with calcitic limestone or other rocks and (*b*) small to large bodies of irregular extent in the midst of calcitic limestones with the separation of the two varieties of limestone not coincident with bedding planes. It is assumed that the units of irregular shape and distribution are of secondary origin. The regularly interbedded units may be of primary deposition or may have undergone dolomitization after deposition but before lithification.

Origin of Dolomitic Limestones

Theories relating to the origin of dolomitic limestones may be placed in the two classes of primary and secondary. Arrangement of the theories is as follows:

- Primary deposition of dolomitic limestones.
  - Deposition as clastics.
  - Deposition by inorganic chemical processes.
  - Deposition by organic chemical processes.
- Secondary deposition of dolomitic limestones.

PRIMARY DEPOSITION OF DOLOMITIC LIMESTONES

Deposition as Clastics. Terranes of dolomitic limestones produce rock fragments as do other kinds of rocks, and the fragments range from clay to gravel dimensions. These fragments may be aggregated and cemented to form a dolomitic limestone composed of clastic particles. The extent to which dolomitic limestones of this origin have been formed is not known, but limited occurrences of dolomitic sands are known. This interpretation has been applied to dolomitic limestones near Harrisburg, Pa., to impure dolomitic limestones in the Muschelkalk of Germany, and to some dolomitic limestones in the Salina of New York.

Deposition by Inorganic Chemical Processes. Fundamental to any chemical theory of the origin of dolomitic limestones is the matter of the relative solubilities of calcium and magnesium compounds (Kline, Wattenberg, Wells). Calcium carbonate is generally considered less soluble than magnesium carbonate under atmospheric pressure and in the presence of carbon dioxide in distilled water, whereas calcium hydroxide is more soluble than magnesium hydroxide. Judd, however, states that under existing atmospheric conditions sea water provided with carbon dioxide dissolves calcium carbonate more rapidly than magnesium carbonate. Skeats confirmed the statement of relative solubilities, but stated that under a pressure of 5 atmospheres the magnesium carbonate in dolomite is dissolved and calcium carbonate is little affected. If this is the case, the two compounds should be equally soluble at some pressure between 1 and 5 atmospheres and then might be precipitated together from saturated solutions. Later studies of the relative solubilities of the two compounds have shown that magnesium carbonate is the more soluble in distilled water. Such might not always be the case in sea water. Goldman found a decrease in the magnesium content of shell materials on the shores of the Bahamas after exposure to sea water, and Leather and Sen state that magnesium bicarbonate is much more soluble in water than calcium bicarbonate, that calcium carbonate becomes practically insoluble in waters containing an excess of magnesium bicarbonate, that a mixing of solutions of calcium and magnesium carbonates leads to precipitation of most of the calcium carbonate, and that if a mixture of calcium and magnesium carbonates is subjected to leaching by water containing carbon dioxide there is little or no solution of the calcium carbonate. Irving has also noted that magnesium carbonate is more soluble than calcium carbonate. Within certain limits, rise in salinity is said to increase the solubility of calcium carbonate.

Wattenberg and Timmermann have shown that the solubility of magnesium carbonate in distilled water is an inverse function of the pH value and Wattenberg stated that magnesium carbonate is less soluble in sea water than in fresh water and that sea water is not saturated with magnesium carbonate, despite the fact that there is about three times as much magnesium as calcium dissolved therein. Precipitation of calcium carbonate is common in sea water, but there is ordinarily little precipitation of magnesium carbonate. It is possible that, under some conditions of pressure and chemical character of sea or lake water, magnesium carbonate may be less soluble and may then be precipitated directly as the carbonate or combined with calcium as the double carbonate. As magnesium hydroxide is more insoluble than calcium hydroxide, the precipitation of the magnesium may be in the form of the hydroxide, which would later change to the carbonate.

Many sedimentary deposits of seas and lakes contain small quantities of magnesium, generally assumed to be in the form of the carbonate. The cause of the deposition is not known.

Dolomite has been found in coral reefs of near Recent age, but the occurrences are clearly the result of secondary processes. The limestones associated with evaporites are very commonly dolomitic. The association suggests that perhaps the constituents of these limestones were deposited as evaporites and that the formation of dolomite might have been contemporaneous with deposition or shortly subsequent thereto. The association has also suggested that high concentration of magnesium may have been a factor concerned in the deposition of dolomite. Bearing on this suggestion is the discovery of Eardley that crystals composed of $CaCO_3 \cdot 2MgCO_3$ have formed and probably are now forming on the bottom of Great Salt Lake, in the waters of which the magnesium content is slightly more than 17 times the calcium (2.76 to 0.16 per cent of the solid matter in solution). These crystals were termed dolomite by Eardley, but they could hardly be this with the stated composition. Their optical properties remain to be determined.

Many students have advanced the theory that primary deposition of dolomitic limestones was brought about by inorganic processes. The Muschelkalk has been assigned to a physicochemical origin, assuming the calcium and magnesium carbonates to have been transported as colloids stabilized by protective organic colloids and finally precipitated in an intimate mixture that subsequently crystallized to dolomite. The Stockbridge limestone of western New England and adjacent New York has been referred to an inorganic chemical origin, the limestones in the Permian of England have had the same reference, and there have been many others. Definite proof, however, has been lacking.

Experiments seeking to form dolomite have not been very successful. It has not always been certain that dolomite was produced, and, as geological conditions were not met, any results have doubtful application. Mitchell seems to have approached most nearly to approximation of geological conditions. He used 500 cc. of artificial sea water in which there was neither magnesium chloride nor calcium carbonate. To this were added solutions of N/25 magnesium chloride and saturated calcium bicarbonate. The solutions were added drop by drop from opposite sides of the beaker containing the artificial sea water until 100 cc. of each had been added. The solution was then made alkaline by adding N/20 sodium carbonate. About 6 hr. later this produced a precipitate which after a week had doubled in bulk. The particles of the precipitate were stated to be too small to permit accurate determination of the optical properties beyond that the refractive index was slightly greater than 1.69 and birefringence was very strong. Analyses of the precipitate gave CaO 26.45 to 26.50, MgO 25.10, and CO_2 48.58 to 48.69 per cent. The content of magnesium oxide is a little too large for dolomite, and it is not certain that any dolomite was precipitated.

Irving has stated that solutions containing magnesium and calcium compounds rapidly precipitate magnesium as the alkalinity rises due to increase of free base and as the pH rises above 10. Precipitation of calcium also takes place, but more slowly. Calcium is precipitated chiefly by increase in carbonate-ion concentration. Thus, if alkalinity in sea water were produced by free carbonate, calcium would exceed magnesium in any precipitate, but if alkalinity were produced by free base the reverse would be true, and, as the sea water approached the pH value of 10, both calcium and magnesium would be precipitated, the calcium as the carbonate, the magnesium as the hydroxide. Wattenberg and Timmermann have shown that if sea water becomes strongly alkaline and the pH rises above 9.5, the solubility product of magnesium hydroxide is exceeded and magnesium hydroxide is precipitated. This might readily be done in concentrated waters from which the carbon dioxide was extracted by plants, thus leaving the water alkaline because of free base. Any magnesium hydroxide precipitated would later change to the carbonate. It is not known that dolomite would form.

The status of the primary precipitation of dolomitic limestones by inorganic chemical processes is thus one of considerable lack of knowledge, and as yet very largely within the realm of theory. It seems probable that the concurrence of a considerable range of various factors is essential.

Primary Deposition by Organic Chemical Processes. The protective and skeletal structures of certain foraminifera, Alcyonaria, crinoids, and some algal structures formed by photosynthesis contain more or less magnesium carbonate, the quantity being greatest in tropical regions (Clarke and Wheeler). The quantity precipitated by some calcareous algae may be as great as 25 per cent of an algal deposit. It is not known that any of this magnesium carbonate is in the form of dolomite. The utilization of carbon dioxide by plants increases the alkalinity and pH, and perhaps leads to precipitation of magnesium hydroxide with calcium carbonate precipitated by photosynthesis. The magnesium hydroxide may be expected to change to magnesium carbonate of which some may return to solution in case alkalinity declines, thus accounting for some of the porosity of many algal deposits.

DOLOMITIC LIMESTONES OF SECONDARY ORIGIN

Many dolomitic limestones are obviously of secondary origin, either by replacement of calcium carbonate by magnesium carbonate, through addition of magnesium carbonate, or leaching of calcium carbonate. It is thought that evidence for all of these processes has been found.

Replacement may have taken place either in the sea before lithification or after the sediments had been lithified and raised above sea level. Replacement before lithification may be over extensive areas. This may have been done for each bed as the bottom was built to a profile of equilibrium and then dolomitized before rise of sea level permitted deposition of another bed. Local conditions in the bottom sediments may produce local replacement. This method of origin has been suggested by many students for dolomitic limestones of formational extent.

Bearing on the matter of replacement are the relations of dolomitic limestones to bedding planes, the mineralogical composition of any contained fossils, and the relations of the particles to each other and to other minerals.

Fossils in many dolomitic limestones have been changed from calcite or aragonite to dolomite. Some have been replaced by silica, and, as these usually have structures well preserved, it is taken to indicate that silicification was antecedent to or contemporaneous with dolomitization.

The crystals of dolomite in limestones containing both dolomite and calcite interlock and are rarely euhedral, thus indicating that the rock at the time of formation possessed sufficient rigidity to prevent crystals from forming without interference. Contacts between the dolomite and calcite crystals are sharp, and there is little or no evidence that the crystals of dolomite obtained calcium carbonate from the calcite crystals that remain.

It has been assumed that the magnesium to form dolomite was derived from sea water, that the magnesium carbonate united with already deposited calcium carbonate to form the dolomite, and that a chemically equivalent quantity of calcium carbonate passed into solution. This would be attended by a 12.3 per cent decrease in volume and would not be likely to take place in normal sea water but might do so in waters of higher concentration of magnesium carbonate such as might be present over shallow bottoms of the sea or over the bottoms of enclosed arms of the sea, particularly if the enclosed arms were in dry regions. It is usually assumed that replacement proceeded according to the following equation:

$$2CaCO_3 + MgCO_3 = CaMg(CO_3)_2 + CaCO_3 \text{ (in solution)}$$

It has also been suggested that addition of magnesium carbonate to calcium carbonate has been responsible for some dolomite occurrences, and actual cases of addition of magnesium carbonate to sediments are stated to have been observed. This leads to increase in volume. Addition seems unlikely in normal sea water, which, as noted, is unsaturated with magnesium carbonate, but it could take place in waters of higher concentration which had been made alkaline by plant action.

Enrichment of sediments with magnesium carbonate could also be brought about by leaching of calcium carbonate, but serious difficulties follow in the train of this process. A sediment with 25 per cent magnesium carbonate and 75 per cent calcium carbonate would be compelled to lose three-fifths of the calcium carbonate before the composition would approximate that of dolomite. If the rock were to some degree lithified, there would be development of large cavities and some collapse of parts of the rock. It might well be done in unlithified sediments without producing collapse or porosity effects. Some sediments clearly give evidence of leaching in the occurrence of fossils as molds. If addition of magnesium and leaching of calcium carbonate proceeded simultaneously, neither collapse nor development of porosity need take place.

Facts showing what actually takes place in dolomitization of calcium carbonate sediments were learned in studies made of samples acquired from a hole drilled in the Funafuti Atoll in the East Indies to a depth of 1,114.5 ft. (339 meters) (Judd). The section exposed in the well showed a moderate quantity of magnesium carbonate to the depth of 35 ft., with evidence of leaching producing concentration of magnesium carbonate at 15 and 25 ft., with 16.4 and 16 per cent of magnesium carbonate, respectively. From 35 to 637 ft. the quantity of magnesium carbonate ranged from 1 to 6.8 per cent, with an average of 3.33 per cent. None of this was known to be dolomite, and the magnesium carbonate could be readily accounted for as due to magnesium carbonate in algal matter. From 638 ft. to the bottom the magnesium carbonate content averaged 40 per cent. From 638 ft., where the magnesium carbonate content rose in a thickness of 2 ft. from 2.44 to 20.44 per cent, to the bottom, the materials of the core were dolomite with some calcite. Calcite was rare or absent from 650 to 820 ft. and from 875 to the bottom. As less calcite was shown in the core samples than indicated by chemical analyses, some calcite is considered to have been in the dolomite crystals. Dolomite was not commonly present if the magnesium carbonate was less than 15 per cent. It is assumed that this quantity could be held in the calcite crystal. Dolomite commonly appeared if the magnesium carbonate exceeded 15 per cent. Aragonite was common in the upper part of the core, it was almost gone at the depth of 150 ft. and none was seen below 220 ft. Aragonite was not seen in association with dolomite. The core indicated some leaching at places in the upper 35 ft., and there may have been some at greater depths. Replacement of calcite or aragonite was proven by the fact of the replacement of shells and corals, and dolomite was found filling original pore space so that there was some addition. Enrichment of the sediments in magnesium carbonate had unquestionably taken place in the upper 35 ft. and to a greater degree from 638 ft. to the bottom. Direct deposition of dolomite is not known to have taken place at any time as the sediments were in process of deposition.

In strong contrast to the Funafuti boring with respect to content of dolomite are those made on Bikini Atoll, Marshall Islands, in the summer of 1947. Five holes were drilled, one to the depth of 2,500 ft. (about 750 meters) and Ladd and Tracy report that little magnesium carbonate was found in the samples analyzed to date of publication (Jan. 16, 1948).

Enrichment of dolomite in calcitic rocks by ordinary ground waters is related to channels of underground circulation, and it also may take place from the surface downward. Occurrences are not known to be numerous.

Formation of dolomitic limestones by hydrothermal replacement is known to have taken place on a large scale. Consideration of these limestones is beyond the scope of this work.

Places of Origin of Dolomitic Limestones of Sedimentary Origin

The fossils of Paleozoic and later dolomitic limestones show that the original sediments from which most originated were deposited under marine conditions. Some were deposited in evaporating basins. Equal confidence can not be expressed with respect to Pre-Cambrian dolomitic limestones, as the only organic structures are noncommittal on the matter of the salinity of the water. Most dolomitic limestones seem to have been deposited in shallow waters, of which some were so shallow that the bottoms at times were exposed and mud cracks developed. Shallow waters favor a higher concentration of magnesium carbonate than deep, and it seems very probable that such concentration is essential for precipitation of magnesium hydroxide or carbonate. High concentration does not seem possible in deep waters. Shallow epicontinental seas surrounded by lands that were not contributing much fresh water and with poor connections to the open sea seem to have been favorable environments for the necessary concentration of magnesium compounds. The rate of deposition of all sediments is thought to have been slow, and it is thought that bottoms near the base level of deposition gave particularly favorable conditions. Shallowness of water would permit repeated stirring of sediments, thus making replacement more readily done. Waters high in magnesium compounds and low in those of calcium would seem to be favorable for replacement of calcium in deposited sediments. Warmth is considered a favorable factor, in that concentration of salts in solution is thus favored, and luxuriant aquatic plant growth over parts of the area of deposition would favor production of alkalinity necessary for precipitation of magnesium hydroxide. It is suggested that the concurrence of these conditions in some basins of deposition and not in others may explain the geographical distribution of dolomite formations.

Summary

The data relating to dolomitic limestones not due to hydrothermal processes permit the following provisional statements:

1. Most dolomitic limestones were deposited in the sea or in salt lakes.
2. Additions of magnesium carbonate, replacement of calcium carbonate by magnesium carbonate, and leaching of calcium carbonate are the only factors relating to the formation of dolomitic limestones that are supported by any considerable body of data.
3. Most replacement, leaching, and addition are considered to have been done in marine waters before the sediments were completely lithified.
4. There is little evidence supporting the view of direct precipitation of dolomite, but it is considered to have taken place. Crystals composed of $CaCO_3 \cdot 2MgCO_3$ are forming on the bottom of Great Salt Lake.
5. Some dolomitic limestones are probably of clastic origin.
6. Most dolomitic limestones were formed in warm shallow waters.
7. The fossil matter in dolomitic limestones does not indicate environmental conditions, at the times these animals lived, different from those in which equivalent calcitic limestones were deposited. It does not follow that the organisms could endure the conditions that may have existed when dolomitization took place.
8. Limestones associated with evaporites are commonly dolomitic and contain few or no fossils.

It is thought that it will be found that dolomitic limestones, like many other sediments, are not all formed according to a single pattern.

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Sedimentary Magnesite

Sedimentary occurrences of magnesite are not common, and, evidently, rare concurrences of certain sedimentary conditions are essential for formation. Magnesium carbonate is present in some form in the shells of many marine invertebrates, particularly those dwelling in the warmer waters of the sea, and some algal structures contain as much as 25 per cent. The quantity in many instances is too large to be in solid solution in calcite, dolomite is not present, and the magnesium carbonate is most likely to be present in the form of small particles of magnesite.

Magnesite deposits near Bissell, California, are in the form of thin-bedded, compact white rock interbedded with gray and green clays (Gale). Deposition seems to have taken place in a lake. An extensive deposit of magnesite is near Overton, Nevada, in the Tertiary Horse Spring formation (Longwell). This formation is underlain by the Overton fanglomerate, and it consists of interbedded dolomite, magnesite, clay, silt, some gypsum, and some tuffaceous beds. The magnesite composes only small parts of the sections, is thin-bedded to more or less massive, and the color is white. Beds are dominantly composed of magnesite, but associated substances in the beds are dolomite, parasepiolite (?), clay, silt, sand, and tuff. Beds are thin-laminated, and have much continuity. Some beds are mud-cracked, some contain crystal molds, and some are composed of (or filled with) pellets referred to excremental origin. These various typical sedimentary features and the absence of any features suggesting replacement led Longwell to postulate that the Horse Spring formation is a playa deposit successive to the Overton fanglomerate of alluvial-fan formation. To account for the magnesium, Longwell postulated that it may have been brought to the playa by hot springs. This does not seem necessary, as the waters of many regions are known to contain considerable quantities of magnesium compounds. Gale suggested for the deposit at Bissell, California, that precipitation was caused by reaction of magnesium sulphate with sodium carbonate. Longwell followed this suggestion. Rubey and Callaghan suggest precipitation of the magnesium carbonate as a consequence of evaporation of lake waters containing much magnesium in solution. Studies of the solubilities of dolomite, parasepiolite (?), and magnesite, the three minerals commonly associated in the Horse Spring magnesium-containing beds, indicated that magnesite is most soluble at low temperatures and dolomite and parasepiolite (?) at high temperatures. This suggested to Rubey and Callaghan that seasonal temperature variations may have played some part in the sequences of sediments shown in the sections. If the three compounds were present in saturated condition in cool waters, warming of the waters would lead to supersaturation and precipitation of magnesite and perhaps some solution of previously deposited dolomite and parasepiolite (?). If evaporation accompanied the warming there might be some precipitation of dolomite and parasepiolite (?). This would give rise to a layer of relatively pure magnesite, or to magnesite containing a greater or less quantity of dolomite and parasepiolite (?). Reversing the change of temperature would lead to precipitation of dolomite and parasepiolite (?), and, if the cooling were due to inflow of cold water, there might be complete solution of previously deposited magnesite and perhaps some solution of previously deposited dolomite and parasepiolite (?), leaving a thin leached, and perhaps to some degree broken, layer of the two last minerals. Alternations of warming and cooling, or cooling and dilution, would yield alternations of pure chemical precipitates of the three minerals, or leached rocks of dolomite and parasepiolite (?) mixed with some clastic sediments. Thickness and purity of individual layers would be due to variation in the factors concerned in precipitation, solution, and additions of magnesium to the waters of the playa lake.

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CHAPTER X

SEDIMENTS OF CHEMICAL DEPOSITION

SILICEOUS, FERRUGINOUS, AND MANGANESE SEDIMENTS

The siliceous, ferruginous, and manganese sediments are associated because of similarity of transportation of the oxides, difficulty of the iron and manganese remaining in the carbonate form under the conditions now existing on the earth's surface, and the seeming probability that somewhat similar environmental conditions are required for precipitation of the three elements in the form of relatively large and pure deposits of the oxides.

Consideration is not limited to the oxides except in the case of the siliceous sediments. All important sedimentary deposits of iron and manganese are considered.

SILICEOUS SEDIMENTS

INTRODUCTION

The siliceous sediments are those dominantly composed of chemically and organically precipitated silicon dioxide. Sediments composed of quartz sands are not included.

The importance of the siliceous sediments cannot fail of appreciation when it is remembered that the stream waters of the world annually carry in "solution" four-sevenths as much silica as calcium carbonate. The total quantity is estimated as 319,170,000 metric tons, or 11.67 per cent of the solid matter in solution; and silicon dioxide holds second place in the quantity of dissolved matter carried, calcium carbonate being first. This silica does not remain in solution for long after the waters reach the sea. Organisms use a small quantity to make shells, but most seems to be precipitated in inorganic ways. The precipitated silica is thus of two kinds: organic, consisting chiefly of the shells of radiolarians and diatoms and spicules of sponges; and inorganic, which includes the siliceous sinters and other spring deposits composed of silica, and flint, chert and jasper.

The shells of radiolarians and diatoms form the radiolarian and diatom oozes. All siliceous deposits are of aqueous origin and most are marine.

Two important and to some extent unanswered problems connected with silica as a sediment are the manner of transportation and the environments and causes of precipitation of the inorganic silica. The source of most sedimentary silica is in the decay of rocks, but some may be contributed to the sea directly by lavas erupted on the bottom of, or in proximity to, the sea, intruded into sediments that form the floor, or by hot springs fed by magmatic sources.

TRANSPORTATION OF SILICA

Most analyses of stream waters show dissolved silica, but studies by Wallace, Baker, and Ward have shown that the "dissolved silica" in the Red River of the North is colloidal and may be precipitated by centrifuge. Moore and Maynard are of the opinion that the silica in natural waters is transported in colloidal form if the concentration does not exceed 25 parts per million, and with higher concentrations a part may be transported as alkaline silicate. Roy concluded that much silica is in true solution as the silicate ion SiO_3^-. The average silica in "solution" does not exceed 15 parts per million. Alkaline waters are more potent than others in removing silica from source rocks, and Lovering found that magnesium and calcium bicarbonates are two of the best solvents. It seems generally agreed that silica in solution is usually carried as alkaline silicate.

SILICA OF ORGANIC ORIGIN

Silica Deposited by Sponges

Some sponges construct spicules of opal or colloidal silica, which on death of the sponges are either deposited on the sea bottom or again enter into solution. The sponges evidently obtain the silica from silica or silicates in "solution" or colloidal suspension. Spicules of hexactinellid sponges are those most common in deep waters but are less common in shallow waters than are those of the tetractinellid and monactinellid sponges. Sponges that form siliceous spicules are sufficiently common locally to make important contributions to bottom sediments, but even in sediments of slow deposition the spicules form only a small part, and in sediments of rapid deposition they are hardly apparent. It may be that many spicules are dissolved not long after the deaths of the sponges that made them. Siliceous spicules are present in the geologic column since the early Paleozoic, but generally they are in isolated individuals. At one time many considered that sponge spicules to a large degree were responsible for the silica of chert and flint. This view does not have much standing at present.

Diatom Ooze and Diatomite

Diatoms are small microscopic members of the plant kingdom that live in both fresh and salt waters. Most belong to the plankton. In the sea they seem to be most abundant in waters of low salinity. They construct shells composed of opal, and accumulations of these shells form *diatom ooze. Diatomite* is indurated ooze. The most extensive deposits of diatom ooze are in the Southern Hemisphere between the Antarctic Circle and 40° south, where 10,880,000 square miles of the sea bottom are covered with this sediment. There is a small area of about 40,000 square miles in the North Pacific. Depth distribution of diatom ooze in the sea is between 100 and 1,975 fathoms. Diatom tests, however, are found on all shallow bottoms but fail to make diatom ooze because of abundance of other sediments. The failure of ooze to accumulate on deep bottoms is due to the shells passing into solution before bottom is reached. Diatom ooze also forms in fresh waters, and most lake deposits contain shells of many species of diatoms.

Diatoms at times die in enormous numbers, as in 1925 at Copalis Beach on the coast of Washington. The so-called epidemic was ascribed to heavy rains followed by westerly winds and much sunshine. A continuous ridge of the diatomaceous matter 10 to 15 cm. in height extended along the beach for 20 miles. It accumulated in a very short time.

Diatom ooze almost invariably contains other organic matter in addition to shells of diatoms, and there is also more or less inorganic matter. Calcium carbonate may range from nothing to a third, and there is invariably some material of terrigenous origin. Color of wet ooze ranges from yellow straw to cream.

Dry ooze is light blue to yellow. The color of diatomite is from light yellow to white.

Diatoms are not known before the Tertiary, and in most cases the shells are not sufficiently abundant to form diatomite. There are some very pure diatomites in the Tertiary of the West Coast area of the United States (Bramlette).

Radiolarian Ooze and Radiolarite

Radiolaria are microscopic animals that have always lived in the sea. Most are plankton and dwell in the surface waters, where they have wide distribution, particularly in the warmer waters. They seem to be most abundant in the central Pacific and eastern Indian oceans. Silica for the shells, which are composed of opal, is obtained from that in "solution" and perhaps from silica and silicates in suspension.

Radiolarian ooze has shells of Radiolaria as the greatest single constituent, but some shells may be seen in most marine sediments. Both organic and inorganic constituents of varied origin usually are present. Manganese nodules, shark teeth, and other resistant parts of vertebrates are occasionally common. Color ranges from gray to red. *Radiolarite* is indurated radiolarian ooze.

Radiolarian ooze as a distinct deposit is largely confined to the Pacific and Indian oceans, with an estimated area of 1,161,000 square miles in the former and 1,129,000 square miles in the latter. The depth range is from 2,350 to 4,475 fathoms, but radiolarian shells may be found in most shallow-water sediments of warm waters, where they fail to make ooze because of abundance of other sediments. Absence from very deep bottoms arises from their passing into solution before bottom is reached.

Radiolaria seem to have been in existence since the Cambrian. The oldest known radiolarite is in the Devonian of New South Wales. This is stated to have a thickness of 2,700 meters and to contain 1 million Radiolaria to 1 cu. in. Radiolarite is present in the Jurassic of the Alps, and with the interbedded shales and sandstones has a thickness of 10 to 25 meters. Radiolarites have been described from the Carboniferous of the Lower Rhine Region and Great Britain, and cherts containing Radiolaria have been described from the Ordovician of New York, the Mississippian of western England, and the Jurassic (Franciscan) of California. The tendency has been to ascribe radiolarites, and diatomites as well, to deep-sea origin, a tendency that does not seem to have much positive supporting evidence.

INORGANIC DEPOSITS OF SILICA

Siliceous Sinter (*Geyserite*)

Siliceous sinter is a deposit of geysers or hot springs with alkaline waters. The silica is a variety of opal, and there is a content of water ranging from about 9 to 13 per cent. A deposit is more or less porous and usually nicely laminated.

The precipitation of the silica in siliceous sinter has been referred to relief of pressure consequent to reaching the surface, cooling, chemical reactions, and algae. Algae are said to precipitate the silica in gelatinous form. Color varies with the temperature of the water and is white in hot water and green in cool. Associated algae seem to give the green color.

After death of the algae the silica becomes soft, acquires a cheeselike texture, and the color changes to gray or white. In some cases the pore space in the silica may become filled with silica of later deposition, and a compact mass is formed. Compared to other inorganic deposits of silica, siliceous sinter has little quantitative significance. Distribution is very local, and thickness and areal extent are both small.

Chert (*Flint, Jasper*)

Introduction. Chert and kindred rocks are among the most common in the geologic column. All are composed of silica and various impurities. Textures are very fine-grained to cryptocrystalline, and to the unaided eye these rocks seem amorphous. Chert includes varieties that have white, gray, and other light colors, and flint those of dark gray to black colors. Jasper is a red variety with color due to ferric oxide. Novaculite is a variety of chert, originally variously interpreted as a fine-grained sandstone or replacement of lime carbonate by silica. It has been shown by Miser and Purdue to be chert. Jasperoid is a name applied to a chert associated with the lead and zinc ores of the Tri-State District, and jaspilite is a form of metamorphic chert associated with Pre-Cambrian iron formations of the Lake Superior Region. It commonly has red color.

These substances are composed of silica and impurities. Common impurities are hematite, limonite, pyrite, marcasite, carbonaceous matter, and calcium and magnesium carbonates.

The silica in the Mesozoic and Paleozoic cherts and flints is crystallized, and a thin section shows a mosaic of quartz and chalcedony. Particles are very small, and in jasperoid from southwestern Missouri the dimensions range from 0.02 to 0.06 mm. Mesozoic and Paleozoic cherts and flints are generally very compact, but many Pre-Cambrian occurrences have been recrystallized to resemble quartzite. Many of the younger cherts and flints contain opaline or amorphous silica. Weathered chert becomes white and chalklike.

Some cherts and flints contain considerable calcium carbonate, and some grade laterally into limestones. The calcium carbonate may be in crystalline particles or pieces of shells.

The color of some chert and flint is uniform throughout, in others the colors are arranged in irregular patches so as to give a mottled appearance, and in still others the colors are arranged in bands concentric to a central area or parallel to the stratification of the enclosing beds. Colors are usually due to some impurity, but they may be due to differences of texture or porosity.

Many cherts and flints are banded, and the banding is due to differences of either color or texture. In some the banding is parallel to bedding, in others concentric to some central area, and in rare cases small patches are banded with the banding of different orientation in different patches. Cherts and flints with concentric banding tend to have the inner bands of darker color than the outer. The bands generally seem to be wider laterally than above or below, and widths range from paper-thin to several centimeters.

Typical chert and flint have dense texture, vitreous to waxy luster, and conchoidal fracture. Some are granular with irregular fracture.

Fossils may or may not be present in chert and flint and are like those in the enclosing strata. Fossils may be distributed throughout or largely confined to outer parts. They may be composed of original shell materials or have been replaced by silica. Preservation is commonly excellent and often much better than in the enclosing strata.

Occurrence of Chert and Flint. Chert and flint occur in beds and lenses and as variously shaped nodules (concretions), cavity fillings, and cement for other sediments.

Nodular cherts and flints are usually associated with limestones. Flints are mostly in chalk and cherts in other varieties of limestone. Bedded cherts and flints have association with all kinds of rocks.

Bedded cherts and flints have extensive distribution in the geologic column. Beds range in thickness from less than 2 cm. to more than 25 cm. The thickness of each bed tends to be uniform and persistent. Well-known formations are the jaspilites of the Lake Superior Pre-Cambrian, the Permian Rex chert of Idaho, which is 18 to 33 meters thick over hundreds of square miles, the Franciscan chert of the Jurassic of California, the Carboniferous chert of North Flintshire, Wales, the novaculite of the Devonian of Arkansas and Texas, and the cherts in the Pennsylvanian of the Mid-Continent region of the United States.

Nodules have all varieties of shapes and are usually arranged along, or parallel to, bedding planes, and surfaces are commonly mammillary. They may be united with each other along a plane, and those of one plane may connect with those of adjacent planes. They may be isolated and irregularly distributed within beds, and may unite with each other to form bodies of great irregularity. Nodules may be so abundant as to form most of a bed. Lenticular nodules tend to have the longer axes parallel to bedding planes. Irregularly shaped nodules may have the longer axes at any angle thereto. Most nodules are less than 25 cm. in diameter, but some are 3 meters or more in diameter parallel to bedding and ½ meter or more thick.

There is usually a sharp contact between nodules and the enclosing rock, but some have no definite boundaries and gradually pass into the enclosing rock. The outer parts of a nodule may be porous and have a chalklike appearance. Nodules may contain irregular patches of limestone, of which some are connected with surrounding limestones; and cherts in the Burlington of Missouri have cracks filled with veinlets from the surrounding limestone. Some nodules have hollow cavities lined with crystals of quartz and other minerals.

Cherts and flints of different horizons are usually different, and in any horizon they may be widely distributed with but little variation in character, as the cherts in the Prairie du Chien formation of the upper Mississippi Valley, the cherts in the Pennsylvanian limestones of the Mid-Continent Region, and elsewhere.

The fossils of some horizons are composed of chert or flint with little or no change taking place in the enclosing rocks, and in other cases the interiors of fossils have been filled with chert or flint.

Chert and flint may serve as cement for other sediments, but occurrences are of limited extent. This is locally the case in the Cretaceous formation of France known as Gaize and locally in a conglomerate at the base of the Cretaceous of Texas. Jasperoid of the Tri-State district cements fragments of jasperoid. Pillow lavas of Notre Dame Bay, Newfoundland, have spaces between the pillows filled with chert (Sampson).

Nodular cherts and flints are present in every geologic system. Well-known occurrences are those in the Ordovician Knox dolomite of the southern Appalachians with 240 to 300 meters of chert, the Prairie du Chien formation of the upper Mississippi Valley, the Niagara limestone of the Great Lakes Region, the Lower Devonian of the Appalachian Region, the Mississippian limestones of the Mississippi Valley, and the Pennsylvanian of the Mid-Continent.

Many limestones contain no chert or flint nodules. They are common in parts of the Prairie du Chien formation of the upper Mississippi Valley but are absent in the Platteville of the same localities. They are rare in the Byron and Mayville formations of the Niagara limestones of Wisconsin but are quite common in the Coral Beds. They are wanting in the limestones of the Mingan and Romaine (dolomitic) formations of the Mingan Islands, and chert and flint are not present in the Ordovician and Silurian limestones of the Anticosti section, and there is none in the Silurian of Gotland. This distribution must have significance.

Precipitation of Silica. Transported silica seems to be mostly in the form of a hydrophyllic colloid that is not very sensitive to electrolytes. Additional stabilization is cared for by protective colloids, of which those of organic origin seem most effective. The silica is precipitated through introduction of electrolytes or colloids of opposite sign, and it is possible that organic stabilizing colloids may be eliminated by bacteria. As silica is usually negative, precipitation is effected by an electrolyte or colloid with positive charge.

Experimental studies of precipitation of colloidal silica are not in complete agreement. Tarr treated sodium silicate solutions of 27.07 parts of silica per million, and solutions of half and twice that quantity, with solutions of sodium chloride and magnesium and potassium sulphates in the proportions in which they are present in sea water, and with solutions of sodium chloride and magnesium sulphate separately. He obtained a heavy precipitate of gelatinous silica in each experiment. He concluded that the coagulation of the colloidal silica was produced through neutralization of the negative charges on the silica by the positively charged ions of sodium, potassium, calcium, and magnesium.

Lovering, using natural sea water, obtained no precipitate from sodium silicate solutions containing 30 parts of silica per million. Tarr suggested that this may have been due to the stabilizing effects of organic colloids in the sea water. He did obtain precipitation from sodium silicate solutions containing 490, 6,000, and 8,000 parts of silica per million. The conclusions were reached that 1 cc. of sea water will precipitate 0.0155 gram of silica from a sodium silicate solution under conditions of high concentration, that precipitation is never complete, that no precipitation occurs if there is less than 36 parts of silica per million, and that there may be failure of precipitation in slightly alkaline solutions containing as much as 275 parts of silica per million, even in the presence of a large quantity of electrolytes.

Moore and Maynard used (*a*) solutions of sodium silicate, (*b*) dialyzed sodium silicate, and (*c*) sodium silicate and hydrochloric acid dialyzed after mixing. The silica in solutions *b* and *c* was in colloidal form and most of *a* was in that form. On treatment with different salts variable results were obtained. It was concluded that calcium carbonate and sea salts are two of the best precipitants from solution *a*, and that sodium chloride and potassium sulphate are not so good. Magnesium sulphate gave little or no precipitation from solution *a* containing 30 parts of silica per million, faint precipitation from this solution containing 60 parts of silica per million, and greater effects with increasing concentration up to the experimental limit of 480 parts of silica per million. In no experiment could all the silica be precipitated, the quantity not precipitated increased with concentration, and ranged from 28 parts per million in 30 parts per million concentration to 104 parts per million in the concentration of 480 parts per million. It seemed quite certain that a part of the precipitate was magnesium silicate. Time was found to be an important factor.

Sodium chloride and sea salt were the most effective precipitants of silica from solution *b*. The quantity of precipitation was related to time, and not all the silica could be precipitated. Both magnesium sulphate and calcium bicarbonate were nearly ineffective. Sea salt, sodium chloride, magnesium sulphate, and magnesium bicarbonate had little effect on solution *c* after 75 days. Furthermore, calcium bicarbonate and magnesium bicarbonate even increased the silica in solution by dissolving it from the glass containers.

A summary of Moore and Maynard's results is as follows. Sodium chloride is ineffective on solutions *a* and *c*, but effective on solution *b*. Sodium bicarbonate is ineffective on solutions *b* and *c* but effective on solution *a*. Moore and Maynard suggest that explanation of the differences may lie in the silica being in different states, perhaps in true solution in some cases or in colloidal form perhaps stabilized by adsorption of various ions in others; or perhaps the silica may have different charges in the different solutions, this finding support in the statement of Hardy that silica is positively charged in acid solutions and negatively in alkaline or feebly acid solutions, thus making the same electrolyte ineffective under one condition and effective under another.

Other studies on the precipitating ability of calcium and magnesium bicarbonates are those of Church, who states that 1 mg. of powdered calcite converts a 1 per cent solution of silica into a gel in 10 min. Cox, Dean, and Gottschalk obtained no results on treatment of a dialyzed mixture of water glass and hydrochloric acid with calcium carbonate, but a strong precipitate on treatment with a calcium carbonate solution saturated with carbon dioxide. Lovering's work showed that magnesium and calcium bicarbonates rarely act as precipitants of silica in nature, and that calcium carbonate precipitates colloidal silica only when the concentration is over 40,000 parts per million, a very rare if not improbable concentration in natural waters.

Many formations contain ferric oxide and silica under such relationships that the ferric oxide must have been precipitated first and the silica later. This is in accord with laboratory experiment. Each is relatively stable in the low concentrations present in land waters, but if "solutions" of the two are mixed, the iron and a part of the silica are precipitated immediately. Moore and Maynard, using concentrations of 20 parts of ferric oxide per million and 60 parts of silica per million, obtained precipitation of most of the iron and about half of the silica. If organic colloids are present, both are stabilized and a longer time is required for precipitation. Moore and Maynard found that 1 cc. of peat solutions containing 92.2 parts of organic matter per million stabilized 200 cc. of solutions containing 10 parts of ferric oxide and 30 parts of silica per million. Or 4.5 parts of organic matter per million are all that are required to stabilize the iron and silica in a solution containing 10 parts of ferric oxide and 30 parts of silica per million, and with greater concentration of organic matter there is greater stability. As the concentration of organic matter in natural waters is often much greater than that of the foregoing experiments, it is obvious that there should be no difficulty in fresh waters in preventing precipitation of ferric iron and silica.

When solutions reach the sea, electrolytes in the sea water precipitate both the iron and the silica, the former first, the silica later, thus giving banding to a deposit—a lower band composed of ferric oxide and an upper composed of silica, or a band of ferric oxide passing seaward into a band of silica. Repeated precipitation would produce a thinly banded ferruginous and siliceous deposit.

It is obvious that experiments relating to silica are not altogether in accord with conditions observed in natural waters. Not all the silica is precipitated in laboratory experiments, whereas little of the silica carried into the sea remains unprecipitated.

Places of Precipitation of Inorganic Silica. The almost immediate precipitation of silica on reaching the sea tends to limit major distribution adjacent to inlets of streams. Under average conditions the silica mingles with clastic sediments, particularly clays and silts, and deposition takes place where stream waters mingle with those of the sea. Under most conditions this is near land, but under some conditions may be considerable distances therefrom, particularly in shallow seas if fresh waters flow outward over salt waters. Under conditions of base-leveled lands on which there is an excellent protective cover of vegetation, few clastic sediments are brought to the sea or into salt-water lakes, and the sediments consist largely of dissolved materials, of which the most important might be silica as soils of such lands might long before have lost any contained carbonates. Deposits not masked by clastic sediments are made. Under conditions of slow deposition, precipitated gels of silica might aggregate to form large bodies. Tarr's studies of the paleogeographical distribution of chert and flint led to the view that these accumulated in shallow epicontinental seas in which at the same time the carbonates of calcium and magnesium were in process of deposition, and he rejected the view, so far as it applied to the cherts studied by him, that deposition was adjacent to the inlets of streams. The view is here held, however, that *contemporaneous deposition largely takes place where fresh and salt waters mingle.* The cherts in the Mid-Continent Pennsylvanian strata are readily explained in this way. Formations without chert and flint are assumed to have been deposited in waters some considerable distances from mouths of streams. Lowlands with an adequate protective cover of vegetation, shallow seas in which deposition of clastics is limited, and waters adjacent to inlets of streams are considered the environments that favor formation of contemporaneous chert and flint. If chert and flint exist in rocks that were deposited far from mouths of streams, it seems best to postulate an origin other than syngenetic or refer the silica to magmatic derivation.

In cases where conditions have permitted determination, it is obvious that deposition of most syngenetic cherts took place in salt waters. However, in the case of the abundant cherts in the Lake Superior Pre-Cambrian formations, proof of saltiness has been wanting, but the trend of opinion is that the waters were marine and also shallow. It does not seem likely that extensive deposits of inorganic silica of terrigenous origin could be deposited in very deep waters. Silica of organic or magmatic origin may be deposited in waters of any depth. It has yet to be shown that silica of organic precipitation ever gave rise to extensive deposits of chert and flint.

The most serious objection to contemporaneous deposition of the silica of cherts and flints is that no gels of the character postulated have ever been observed. The answer is that studies of marine sediments are very few, and the physiographic conditions on existing lands are such that streams bring very large quantities of clastics to the sea. The rapid deposition of these clastics precludes formation of globular masses of silica and masks that which is deposited. However, masses of gel-like sediments assumed to be high in silica gel are in process of deposition about the mouth of the Mississippi (Russell).

Time of Origin of Chert and Flint. Hutton referred cherts and flints to intrusive and hence epigenetic origin. For a long time they were considered to result from solution and reprecipitation of organic silica, but the present trend of opinion considers that much chert and flint were deposited contemporaneously with the enclosing rocks. The views of origin may be placed in three classes of contemporaneous (syngenetic), penecontemporaneous, and subsequent (epigenetic). It is considered probable that chert and flint have formed in each of the three ways.

Syngenetic Chert and Flint. Organic precipitation of silica assumes the silica to have been in the form of shells. These are assumed to undergo partial solution after deposition, and the dissolved silica is then assumed to cement the shells not yet dissolved. As evidence for this view there has been offered the presence of sponge spicules and other siliceous organic matter in chert and flint. Perhaps the chief exponent of this view was Hinde, but it has received support from many eminent geologists, among whom are Van Hise, A. Geikie, and Sollas. The theory fails to explain why the spicules and shells in the chert and flint are ordinarily extremely perfect, whereas, if the theory has validity, they should be much corroded by solution. The spicules and shells are more readily explained as having fallen into inorganically precipitated silica in process of accumulation, and hence the presence is incidental and not genetic. In somewhat modified form, the theory assumes that silica derived from solution of organic silica collected in depressions on the bottoms of seas or lakes.

Inorganic precipitation of silica was first proposed in 1888 by Prestwich but did not obtain much support until recent years. Tarr has been one of the strongest of recent advocates. This theory postulates that silica in "solution" is precipitated as a gel by electrolytes in solution in sea water or through bacterial elimination of stabilizing organic colloids. The gel settles to the bottom and assumes spheroidal or ellipsoidal form. If a large quantity is precipitated, a lens or bed is formed. Concentric banding in nodules and lamination in beds represent successive periods of deposition. The gel bodies increase in dimension by direct addition from sea water, and they may enlarge after complete or partial burial by replacement or displacement of adjacent sediments. If deposition of other sediments is slow, growth may be merely enlargement of a nodule; but if deposition of other sediments is rapid, growth of a nodule may continue on uncovered parts, which would then become protuberances. A slowing up of deposition of other sediments may lead to a protuberance expanding into another nodule at a higher stratigraphic level. Protuberances may also be formed by more rapid growth of parts of a nodule. Shells and other sediments from time to time fall into the gel masses, thus explaining the common occurrences of such in chert and flint. Shells and other sediments may also become enclosed in the gel masses if these should roll about on shallow bottoms. The rolling may also give rise to concentric banding. As gels contain much water, shrinkage and cracking are likely as the water is expelled, and surrounding sediments may then penetrate the nodules. The fact that fossils in cherts and flints are often better preserved than in the enclosing rock is hardly explainable on any other basis than that they dropped into accumulating gels. Accumulating overburden flattens spheroidal and ellipsoidal bodies and changes shapes to lenticular. Dehydration and crystallization take place in course of time, and the gels change to chert or flint.

Silica may also be brought to sites of deposition by springs of magmatic origin or by direct contribution from magmas. The temperatures of the waters of deposition may then be above normal. Appeal has been made to this origin for silica to form the immense deposits of chert in the Pre-Cambrian of the Lake Superior Region (Van Hise and Leith), the cherts of the Flintshire Region of North Wales (Sargent) and the Franciscan cherts of California (Davis, Taliaferro). However, Gruner and, with certain qualifications, Moore and Maynard have presented evidence that the silica necessary for the Pre-Cambrian cherts of the Lake Superior Region could have been provided by normal processes of weathering.

Penecontemporaneous Chert and Flint. The theory of penecontemporaneous origin postulates that the silica is precipitated with other sediments. In clastic sediments it remains disseminated, but in carbonate sediments it is assumed to go into solution under certain conditions, later to be reprecipitated around centers, displacing or replacing other sediments, thus eventually building the nodules that become chert and flint. All this is assumed to take place while the sediments are soft. It has not been explained why the silica should again go into solution after precipitation and later be reprecipitated, but the fact that fossils have been replaced by silica after they seem to have been buried in soft sediments indicates that processes something as indicated do take place, and it may be that nodules are formed in the way postulated. Bedded cherts could hardly be formed in this way. A phase of the origin of penecontemporaneous chert and flint may be connected with the formation of dolomite. The dolomitization of soft calcium carbonate sediments may lead to redistribution of the silica and the formation of the chert and flint nodules. This theory of origin has received support from many geologists.

Epigenetic Chert and Flint. The theory of subsequent origin of chert and flint requires that these substances either fill cavities or replace or displace the enclosing solid rock. Filling of cavities with chert and flint does not often take place and displacement is probably rare. Replacement has generally been assumed to be common and to be done by cold ground waters containing dissolved or colloidal silica. For purposes of discussion the regions of replacement may be placed in the two general classes of (*a*) the zone of cementation below the water table and (*b*) the zone of weathering above the water table.

Formation of chert and flint below the water table should be related to such divisional openings as bedding planes, joints, and faults, and to porous rocks. There is little evidence that the cherts and flints in the geologic column are connected with such openings, and veins of chert do not seem to be common. Cementation of porous rocks by silica is common, but the silica is usually in the form of quartz and, rarely, chalcedony. The distribution of chert along bedding planes has been offered as evidence of epigenetic origin, but the chert in these positions is more easily explained as syngenetic. There must also be a source for the silica of epigenetic chert and flint. Removal of silica could not readily be obscured, and there is little evidence therefor. It is difficult to explain how water could extract silica from fine-grained rocks of low permeability, transport it through these rocks, leave no evidence of the extraction, and deposit it along a near-by bedding plane. In many cases the quantity of silica necessary to form the chert and flint that are present is so great that rocks after extraction would be cavernous. There is no evidence to this effect. Many limestones contain beds of chert and flint with extents of hundreds of miles, whereas other horizons in the same formation contain none, although apparently similar in characteristics. This is difficult to explain as a consequence of ground-water action. Chert and flint are frequently found in the middle of beds with no place visible through which water could have entered to bring the silica and remove the carbonate.

The fundamental objections against general application of this view of origin of chert and flint are that (*a*) these rarely show the structural and textural features of the rocks assumed to have been replaced; (*b*) chert and flint nodules seem to bear little or no relationship to channels of underground water circulation; (*c*) fossils are frequently better preserved in the nodules than in the enclosing rocks; (*d*) no immediate source for the silica is obvious, and there is no evidence for removal of silica; and lastly (*e*) there is great difficulty, if not impossibility, of transportation of silica through rock of low permeability.

Formation of flint and chert in the zone of weathering above the water table has more in its support. Weathering of rocks releases silica, and the removal of silica from soils and deposition elsewhere are established facts. The silica may travel laterally or downward, and deposition takes place where precipitants are encountered. Ulrich seems to have been the first who suggested this method of origin, and he explained the cherts in the Lower Paleozoic of Missouri in this way. Robinson pointed out that silicification develops in underlying rock through weathering of overlying rock. In a region that has undergone long weathering with the waters of a character to dissolve the silica it seems likely that much replacement of soluble rocks on the surface of weathering may take place. The view is held that much chert and flint form in this way, and good evidence therefor is the occurrence of fossils of which the exposed parts are silicified, but the parts buried in the rock are carbonates. Van Hise and Leith have suggested that some of the Pre-Cambrian cherts may have been formed through alteration of greenalite as indicated by the following equation:

$$4Fe(Mg)SiO_3 \cdot nH_2O = 2Fe_2O_3 \cdot nH_2O + 4SiO_2$$

Summary. It seems probable that chert and flint are formed in all the ways postulated, but it is considered that very little owes its origin to organic precipitation. It is thought that large bodies of chert and flint represent original inorganic precipitation of silica adjacent to inlets of streams or springs into the sea. Many small bodies have the same origin. Small bodies of flint and chert and replaced fossils within rocks may represent flint and chert of penecontemporaneous formation. Some chert and flint may have formed through replacement of indurated rock below the water table, but most of this is thought to have taken place in the zone of weathering above the water table.

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SEDIMENTARY DEPOSITS OF IRON-BEARING MINERALS

INTRODUCTION

There are few sediments that do not contain iron-bearing compounds, but also few in which these are present in any degree of abundance. Common compounds that contain considerable iron are the carbonates, oxides, hydroxides, and sulphides of iron and certain silicates.

Considerable difficulty has been had with respect to the origin of many relatively pure bodies of iron minerals, but it would seem that there should be little, if any, more difficulty in understanding the origin of these than of relatively pure deposits of limestone. The origin of the latter, however, was made apparently easy by the occurrence in some of large quantities of shells and because of the wide distribution of calcium and magnesium carbonates in solution. Iron, however, is a more abundant element in the earth's crust than either calcium or magnesium, and under existing conditions it is known to be removed from many residual materials derived from decay of iron-bearing rocks.

The problems connected with iron in sedimentary deposits center around the three heads of (*a*) methods by which iron is transported from places of release from parent rocks to places of deposition, (*b*) methods by which deposition is accomplished, and (*c*) environmental conditions that made relatively pure deposits of iron-bearing minerals possible. The first two problems are fairly simple. The third presents difficulties for some deposits.

Iron minerals may be released from parent rocks on weathering as little-changed particles and as products of decomposition of original iron-bearing minerals.

Common minerals released without much change are magnetite and ilmenite. The products of decomposition are commonly oxides, hydroxides, and the carbonate. Oxides and hydroxides are relatively insoluble and tend to remain near places of production, but under conditions of an abundance of organic matter they may be removed in colloidal form or be reduced to the ferrous condition and removed in solution.

TRANSPORTATION OF IRON COMPOUNDS

Some particles of iron minerals are transported by traction and visible suspension. These may undergo little change in transportation, or they may experience little to much abrasion and decomposition. The particles travel and are deposited in association with other minerals of approximately equal dimensions or specific gravities.

Gruner's and Moore and Maynard's studies have indicated that under existing conditions most transportation of iron and also most silica is in the form of colloids. The former states that peat solutions are effective aids in transportation. Long transportation requires stabilization by protective colloids, which are generally thought to be largely organic. Moore and Maynard have shown that 36 parts of ferric oxide per million can be transported in colloidal form in waters containing 16 parts of organic matter per million. According to Gruner, the Amazon with an iron content of only 3 parts per million is competent to transport to the sites of deposition 1,940,000 million metric tons of iron in 176,000 years, a quantity approximately equal to that in the Huronian Biwabic formation. It is obvious that no great demand is placed on aqueous transportation to move large quantities of iron.

Some iron is carried in solution as ferrous sulphate, and, locally in regions in which sulphuric acid forms through oxidation of sulphides, the quantity may be large. Some iron is also carried as ferrous chloride.

Considerable has been written respecting the presence in solution of organic salts of iron, but there is little definite information respecting kinds and concentrations. It is probable that iron salts of the organic acids in the form of acetates, malates, citrates, etc., are transported, but that quantities are small.

Considerable iron is probably transported as the bicarbonate. This requires waters with an excess of carbon dioxide and a deficiency of oxygen, and it also generally requires organic matter to bring about reduction of ferric oxide. Ferric oxide in the presence of decaying organic matter is reduced to the bivalent form, which then unites with the carbon dioxide produced by organic decay. Residual materials of moist regions with an abundance of decaying organic matter ultimately lose all contained iron, and most is probably removed as bicarbonate. Under existing conditions iron bicarbonate cannot be maintained for a long time in solution, as oxygen enters the waters and the iron is oxidized and hydrolyzed to ferric hydroxide.

As free oxygen does not seem to be present in volcanic emanations, or in the cellules and smaller openings of igneous rocks and meteorites, and as carbon dioxide is said to be present under both conditions, it has been suggested that early atmospheres of the earth may have been very largely composed of carbon dioxide. As supporting evidence for the scarcity of oxygen in the early atmosphere MacGregor has called attention to the low degree of oxidation of the Rhodesian early Pre-Cambrian rocks. In an atmosphere largely composed of carbon dioxide, all waters would have been highly charged with this gas, and there would then have been little or no difficulty of transportation of iron bicarbonate in large quantities and for long distances. Carbonate of iron would have been the most common product of iron in the weathered materials and in solution, and ferric oxide would have been rare. There would also have been immense quantities of the bicarbonates of calcium and magnesium in solution.

The river and lake waters of North America contain iron in "solution" in quantities ranging from a mere trace to 3.33 per cent of the solids in "solution." The average iron in "solution" in the river waters of the world is 2.75 per cent of the "dissolved" solids. There is little iron in solution in the waters of the sea, and that brought by streams must be early precipitated. Cooper states that the quantity of ferric and ferrous salts in solution in sea water is probably less than 2 mg. per cubic meter and that the concentration of free ferric iron in sea water is not likely to exceed 10 to 12 mg. per cubic meter.

PRECIPITATION OF IRON COMPOUNDS

Iron is precipitated as ferric oxides and hydroxides, ferrous carbonate, hydrous ferric and ferrous silicates, ferrous phosphates, ferric sulphate, and ferrous sulphides. Iron phosphates occur only as occasional particles in other deposits and are not further considered here. Iron sulphate has similar distribution, but it is also formed in limited occurrences about springs, particularly those of mines. Precipitation of iron compounds may be done through intermediation of organisms or through inorganic processes.

Iron minerals transported by traction and visible suspension are deposited as competencies and capacities of the transporting agencies decline with decrease in velocity of the transporting mediums, and the iron minerals then mingle with the sedimentary particles with which they travel. Only rarely may they compose a very large part of a deposit.

Iron transported as chloride and sulphate may ultimately change to carbonate, silicate, oxide, or hydroxide. Both hematite and magnetite are said by Van Hise and Leith to develop in this way, as indicated by the following equations:

$$6FeSO_4 + 3O = 2Fe_2(SO_4)_3 + Fe_2O_3$$
$$9FeSO_4 + 4O = 3Fe_2(SO_4)_3 + Fe_3O_4$$

Iron in the form of the bicarbonate under existing conditions may be carried only in waters low in oxygen and containing much carbon dioxide. It may be precipitated as the carbonate if the carbon dioxide is taken from the waters without replacement by oxygen. This may be done by bacteria utilizing the carbon dioxide of the bicarbonate for food or by green plants in photosynthesis. Green plants at the same time release oxygen, but the quantity is small in comparison to the carbon dioxide utilized. Under conditions of a high carbon dioxide content in the atmosphere, as has been suggested by Barrell, MacGregor and others, there would be difficulty of precipitation of carbonate. Calcium and magnesium bicarbonates are more soluble than iron bicarbonate, and if precipitation were possible, the iron salt would leave the water in advance of those of calcium and magnesium. These would be precipitated at a later time and probably at much greater distances from the places of entry into the basins of deposition.

Under existing conditions, iron transported as the bicarbonate is very likely to be changed in transportation to the oxide or hydroxide, bacteria being in some degree responsible for the change. This is shown in the following equations:

$$FeCO_3 + H_2O = Fe(OH)_2 + CO_2$$
$$4Fe(OH)_2 + 2O = 2Fe_2O_3{\cdot}3H_2O + H_2O$$

Precipitation of iron compounds related to bacterial or other action is common in springs and bogs, but it may take place anywhere if oxygen enters the waters and carbon dioxide can escape. It may be assumed that similar precipitation took place in the past and that some of the extensive deposits of iron oxides in the geologic column may have originated in this way. There is little evidence, however, of bacterial action in the ancient deposits, which is not difficult to understand if one takes into consideration the fragile nature of bacteria. Structures identified as due to bacteria have been observed by Walcott and Gruner in Pre-Cambrian rocks and by Bleicher in the Minette ores of France. The relation of the precipitation of iron salts to bacteria (Baier, Harder) has not been completely solved, and three possibilities have been suggested, namely, (*a*) the iron-precipitating organisms require the iron compounds in their life processes much as the sulphur bacteria utilize sulphur compounds, or nitrogen-fixing bacteria utilize nitrogenous compounds; (*b*) the organisms use the iron compounds simply for the purpose of external protection, as many organisms use silica and calcium carbonate; and (*c*) the ferric hydroxide is due to chemical action and incidentally collects on the organisms because of the coverings of mucilaginous matter.

It has been shown by Harder that each of these possibilities, singly or severally, probably applies to some group of bacteria. Studies, however, indicate that most precipitation of ferric hydroxide by bacteria is due to life processes of the organisms. Certain bacteria, as the common *Spirophyllum ferrugineum* and possibly *Gallionella ferruginea*, precipitate ferric hydroxide from ferrous carbonate and use the carbon dioxide liberated and the energy produced in the oxidation. These bacteria cannot exist without ferrous carbonate in solution. There are other "thread bacteria," as *Leptothrix ochracea*, *Cladothrix dichotoma*, and probably others, that do not require ferrous carbonate in their life processes but do precipitate ferric oxide if lime salts are in solution. *Leptothrix ochracea* is abundant in iron-bearing spring and marsh waters. A third group of bacteria requires organic salts of iron in solution and does not use inorganic salts of iron. These salts are attacked with the probability that the organic radical is used for food and ferric hydroxide is ultimately precipitated. The bacteria of this group seem to be the least well known.

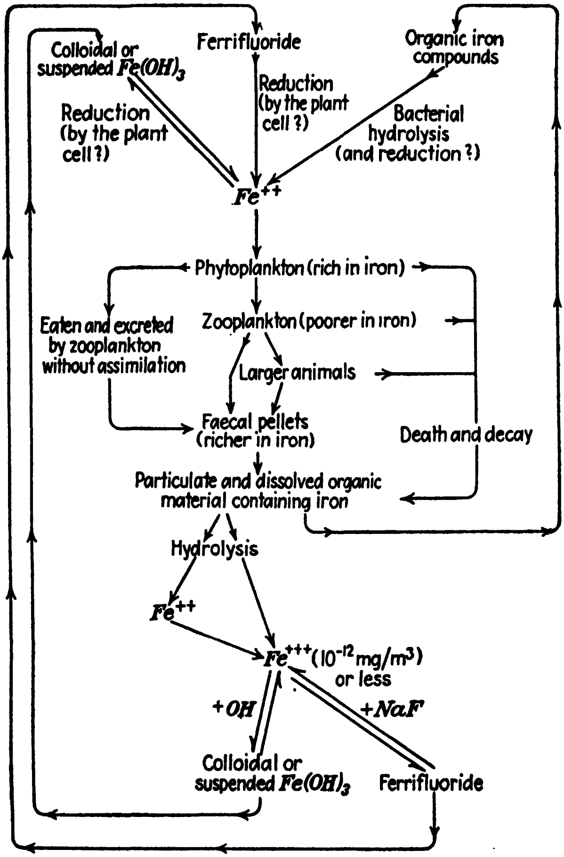
Colloidal or suspended Fe(OH)3
Ferrifluoride
Organic iron compounds
Reduction (by the plant cell?)
Reduction (by the plant cell?)
Bacterial hydrolysis (and reduction?)
Fe++
Phytoplankton (rich in iron)
Eaten and excreted by zooplankton without assimilation
Zooplankton (poorer in iron)
Larger animals
Faecal pellets (richer in iron)
Death and decay
Particulate and dissolved organic material containing iron
Hydrolysis
Fe++
Fe+++ (10-12 mg/m3) or less
+OH
+NaF
Colloidal or suspended Fe(OH)3
Ferrifluoride

FIG. 45. Tentative diagram of the iron cycle in the sea. [*After L. H. M. Cooper, Iron in the sea and marine plankton, Proc. Biol. Sci., Ser. B*, 118 (1935), *Fig.* 2.]

Iron-precipitating bacteria seem to be present wherever waters contain salts of iron in solution. The quantity need not be large. The organic matter in solution affects distribution of iron-precipitating bacteria, some requiring much, others preferring waters with little or none. Not a great deal is known respecting iron-precipitating bacteria in salt waters. They have been identified in brackish and salt-water lagoons on the Texas Coast (North and Bridenstine) and in White and Petchora seas (Butkevich), and it may be that distribution is general in the sea and that varieties and species are adapted to every marine environmental condition. Temperature within viable limits seems to have little effect, but most forms studied seem to prefer waters in the range from cool to warm. Lieske gives the optimum temperature for *Spirophyllum* as 5 to 6°C., and Harder found that a common temperature in "iron springs" containing iron-precipitating bacteria is between 9 and 10°C.

Iron is an essential element for many organisms, and diatoms are said by Cooper to require several times as much iron as phosphorus. Much of this iron remains permanently in organic circulation, but some is deposited with burial of the organic matter and in part becomes responsible for the iron generally present in organic deposits. The iron cycle with respect to organisms is shown in Fig. 45. The quantities of iron used and precipitated by organisms are not known.

Environmental Conditions Making Possible Relatively Pure Deposits of Iron Minerals

INTRODUCTION

The most difficult problem connected with some iron-bearing sediments relates to the environmental conditions that make possible the deposition of large bodies of iron-bearing minerals relatively free from contamination by other substances. Dissemination of iron-bearing minerals in other substances offers little or no difficulty, as there are many ways by which this may be brought. Deposits of iron-bearing minerals have been formed under conditions of a wide range of environments that include those of weathering, seeps, springs, swamps, streams, lakes, lagoons, and the sea. Deposits connected with weathering, seeps, springs, marshes, small lakes, and streams are not difficult to explain, but the large deposits formed in large bodies of water, either lakes or the sea, present problems that have not been completely solved. Some iron compounds, as ferric hydroxide, are known to be deposited in every environment, whereas others, as the silicates and sulphides, attain best development in the marine environment.

The problem of forming pure deposits is found in the difficulty of separation of the iron compounds from the other sediments with which they are associated in transportation. Large bodies of iron-bearing minerals are interbedded with clays, silts, and sands, and very frequently they are associated with much silica, and it seems obvious that, at least at times, they were associated with these sediments in transportation. The problems are considered in some detail in connection with each particular type of deposit.

The iron-bearing sedimentary deposits are considered in succeeding topics as iron oxides and hydroxides, iron carbonate, iron sulphides, and iron silicates.

IRON OXIDES AND HYDROXIDES

The oxides and hydroxides of iron that are known to be formed by sedimentary processes are hematite (Fe_2O_3), magnetite (Fe_3O_4), turgite ($2Fe_2O_3 \cdot H_2O$), göthite ($Fe_2O_3 \cdot H_2O$), limonite ($2Fe_2O_3 \cdot 3H_2O$), zanthosiderite ($Fe_2O_3 \cdot 2H_2O$), and limnite ($Fe_2O_3 \cdot 3H_2O$). Some iron-containing minerals are derived directly from parent rocks. These are deposited in the same way as other clastic sediments.

Studies by Posnjak and Merwin indicate that most limonite is göthite with adsorbed and capillary water and that turgite is a combination of hematite and göthite with adsorbed and capillary water. Göthite is the only hydroxide that is known to form crystals. Hematite is the most abundant oxide in sedimentary deposits, and the limonite seems to be more abundant than other hydroxides.

The common oxides and hydroxides have the iron in trivalent form, but magnetite with bivalent iron may form through oxidation of ferrous sulphate and probably also as a consequence of bacterial decay of organic matter. There is some magnetite in the Cuban laterites, which is believed to be due to organic action, and Berz has pointed out that magnetite is not uncommon in marine sediments.

Sedimentary oxides and hydroxides of iron are known to have been deposited under residual conditions following decomposition of a wide variety of rocks at one extreme and under marine conditions at the other.

In most cases the oxides and hydroxides of iron are mingled in greater or less abundance with other sediments, of which the most abundant are particles of quartz, clay minerals, and the carbonates of calcium and magnesium. These particles may be more or less uniformly distributed with the iron minerals or interbedded with them.

The iron in the stratified deposits of ferric oxides and hydroxides is considered to have been transported in all the ways possible, but it is considered probable that most of it was transported under atmospheric conditions like those at present as colloids stabilized by protective organic colloids. Precipitation probably took place in several ways, with variations in time and place.

Mechanical Deposits of Iron Oxides and Hydroxides. Rocks that are destroyed without much modification of the contained minerals may have the minerals of low specific gravity moved in advance of those of high specific gravity, or they may be moved together and the heavy minerals later concentrated by wave and current action in stream channels and beaches. The heavy minerals consist largely of magnetite, ilmenite, and hematite, and with these there tend to be associated such minerals as garnet, rutile, zircon, chromite, and the amphiboles and pyroxenes. The deposits may have considerable local importance, and on beaches they are generally known as "black sands." They have considerable local distribution on beaches of Oregon (Twenhofel), Washington, northern California, north shore of the Gulf of St. Lawrence, shores of New Zealand, and elsewhere. Mechanical deposits of hematite in the vicinity of the rich iron deposits in the Pre-Cambrian of Minas Geraes of Brazil are known as *sand and rubble ore* and *canga.* Low-level laterites represent lateritic materials that in some instances have been derived from laterites of higher elevation.

Most mechanical deposits of iron minerals are of recent origin, but there can be little doubt that similar deposits have been made in the past. Few are known. Moderately consolidated deposits of titaniferous magnetite sands have been described from the Upper Cretaceous of northwestern Montana, and it has been suggested that some of the Ordovician siderites and limonites near Hudson, New York, were originally magnetite sands.

Residual Deposits of Iron Oxide. Under certain conditions of weathering, clay minerals decompose to silica and alumina, and the silica is removed from the residual materials. Aluminum and iron hydroxides remain to form laterite, which may range in composition from one extreme of nearly pure iron hydroxide to the other of nearly pure aluminum hydroxide. Some titanium and manganese oxides tend to be commonly present. A profile of a residual lateritic deposit is given by Lacroix as follows:

TABLE 23

	Meters
1. "Iron crust"	0.3 to 2.4
2. Concretionary zone of aluminum and iron hydroxides with oölitic and pisolitic structure	0.9 to 21
3. Bleached zone of light-colored aluminum hydroxide. Retains structures and textures of original rocks	4.5 to 24
4. Passage into original rock	

Laterization is confined to tropical and subtropical conditions. It has been referred to biological processes (Holland), to tropical heat and vegetation under conditions of alternating dry and wet weather (Maclaren), and to maintenance of open texture in clays because of absence of frost action (Mead).

Laterites have wide distribution in well-drained areas of tropical and subtropical regions. The most important and best-known deposits of lateritic iron oxides and hydroxides are those of Cuba. These cover small to large plateaulike areas and range to about 15 meters in thickness. Some aluminum hydroxides are associated. Some magnetite is present, for which an organic origin has been suggested. The deposits are residual after serpentine except in the Camaguey district, where the parent rock is limestone.

Bog and Lake Deposits of Iron Hydroxide. Bog deposits of iron consist of yellow to dark-brown and red mixtures of ferric hydroxide, iron silicate, iron sulphide, iron carbonate, organic matter, clay, and sand. These are known as *murram* in the tropics of Africa and *mocarrero* in Cuba. There is usually a high phosphate content, and wad or bog manganese is a common constituent. If they contain much organic matter, the iron carbonate content is also high. The deposits are in the form of disklike or irregularly shaped bodies, and they form either on the bottom or beneath pervious sediments that overlie impervious sediments. The conditions are generally those of impeded drainage, and in some low areas they take the form of *iron pan*, a variety of hardpan, the *Ortstein* of the Germans.

Iron deposits of lakes are about the margins, and few are said to be in waters deeper than about 0.6 meter, but deposits in the form of crusts at least 15 cm. thick composed of iron and manganese oxides and organic matter cover parts of the bottom of Trout Lake in northern Wisconsin in depths of water of 62 ft.

Many deposits in lakes are concretionary bodies that range to about 0.33 meter in diameter and have internal structure concentric to the nuclei around which they formed. Others consist of light iron-bearing mud filled to a greater or less degree with vegetable matter. This in time hardens into a crust with a blue, black, or green color.

Clinton Type of Iron Oxide Deposits. The Clinton type of iron oxide has wide distribution in the geologic column and is known in many parts of the world. It is represented by the Wabana hematites of Bell Island in Newfoundland (Lower Ordovician), the Neda hematites of eastern Wisconsin (Upper Ordovician), the Brassfield hematites of eastern Kentucky (very low Silurian), the Clinton hematites of the Appalachian Region (Silurian), the mid-Devonian hematites of eastern Kentucky, and the Minette ores of Alsace-Lorraine (Base of Middle Jurassic). There are similar deposits elsewhere in the column, but the bodies are small and rarely have commercial importance.

The Clinton strata of the Appalachian Region are of marine origin. Marine fossils are more or less common to abundant, and in some cases the fossils are composed of hematite. The beds containing hematite extend from New York to Alabama, and the ore beds range in thickness from a few centimeters to 12 meters. The hematite is in the three forms of oölitic, powdery, and replaced fossils. Pebblelike bodies composed of hematite, or of hematite mixed with clay, occur locally near the base. These have a shiny surface like that of desert varnish, but no connection with desert conditions is indicated. The ores are interbedded with clays and shales, and a shallow-water environment is indicated by all the available facts. The fossil ores are largely replaced shells and tests of bryozoans and brachiopods, and the oölitic ores may originally have been composed of calcium carbonate.

Three views have been held respecting the origin of the Clinton hematites. Some have considered the powdery and oölitic ores the form in which the iron oxide was originally deposited. Others consider the oölitic and powdery ores replacements that, according to one school, took place shortly after deposition, or, according to another, long after deposition. It is considered that the latter view of replacement cannot well be maintained, although it may have local application. Perhaps a combination of the view of original deposition and replacement shortly after deposition best meets the conditions to be explained. The fossil ores are certainly replacements.

The Minette limonites are in the form of lenticular bodies interbedded with shales, calcareous strata, and sandstones. The units range in thickness to 6 meters and extend through a vertical thickness of 20 to 45 meters of the associated strata. The limonites have oölitic texture with the oölites of the dimensions of 0.5 mm. or less. The oölites are composed of ferric hydroxide and silica and have nuclei that vary from organic fragments to mineral particles. Cementing materials of the oölites consist of silica, carbonate, or clay. It has generally been considered that the limonites are of primary origin, but Cayeux and a few others have considered them replacements.

The Wabana hematities have oölitic texture with the oölites ranging in diameter from 0.1 to 0.5 mm. They are ordinarily composed of alternate laminae of hematite and chamoisite, but in some cases an oölite is composed entirely of either hematite or chamoisite. Beds range in thickness to 90 meters. The oölites are generally cemented with chamoisite, but in some cases the cement consists of more or less hematite, quartz, or siderite. Calcium carbonate is wanting or rare, but there is much calcium phosphate. Marine fossils are present.

Similar deposits of iron oxides or hydroxides are in the Jurassic of England.

Pre-Cambrian Hematites. The hematites of the Lake Superior Region are typical of the Pre-Cambrian hematites. The original states of much of the iron sediments are postulated to have been in the form of iron silicate or cherty iron carbonate, and the present state of the hematites is postulated to be secondary. Much chert, iron silicate, and iron carbonate are associated, and the iron deposits are interbedded to a greater or less extent with arenaceous and argillaceous sediments. The sediments are obviously of aqueous origin, and deposition seems to have taken place in shallow water. The associated strata have usually been interpreted as having been deposited rapidly, and by implication the iron minerals have also been so interpreted. The view of rapid deposition seems to have been based on the clastic nature of the associated sediments and on the presence of ripple marks and cross-lamination, but not one, or the aggregate, of these characters justifies the view of rapid deposition; certainly any hypothesis of rapid deposition for the clastic sediments need have no application to the sediments of iron minerals and the silica. The quantities of iron and silica have been thought to be too large to permit the view that the iron could have been released from older rocks in the normal processes of weathering or transported by streams to the places where they were deposited. It has generally been assumed that the quantity of organic matter in the Pre-Cambrian was limited, and thus there has been difficulty in understanding how the iron could have been transported as the bicarbonate, or how both iron and silica could have been transported as colloids stabilized by organic colloids. Difficulty of transportation of the iron as the bicarbonate vanishes, if the carbon dioxide content of the atmosphere was high, as has been postulated by some students, and the problems of transportation as colloids stabilized by organic colloids disappear if there was an abundance of organic matter. The extent of graphitic and carbonaceous matter in the Pre-Cambrian rocks, the high degree of mature decomposition of many Pre-Cambrian sediments, the actual occurrence of algal deposits in the Pre-Cambrian strata of many parts of the world, and the actual occurrence of Pre-Cambrian fossil invertebrates in Montana indicate an abundance of organic matter, and thus no difficulty of transportation of colloidal iron oxide and silica. However, it has been assumed that deposition was rapid, and a source for the iron and silica has been sought in magmatic waters derived from lavas, from magmatic waters derived from hot springs, or from reactions of marine waters with lava flows. The magmatic waters or the lava flows are assumed to have carried large quantities of iron and silica, and on contact with the colder salt waters the iron and silica were precipitated as iron carbonate, perhaps iron oxides, iron silicate, and chert (Van Hise and Leith). However, if the atmosphere had its present composition in Pre-Cambrian times, it seems unlikely that iron would have been precipitated as the carbonate if deposits made from the waters of submarine hot springs in the Aegean Sea southeast of Greece in the vicinity of the volcano Santorin are representative of what takes place. The waters of these springs contain iron carbonate in solution, discharge is known to have begun as early as 1650 and to have been practically continuous from 1707 to 1848 and from 1869 to 1933. The sea in the vicinity of the springs is said to be colored yellow-green and red over a considerable area. But the carbonate quickly oxidizes, the color of the water changes to some shade of red, and the precipitates over the bottom are not carbonate, but iron oxide or hydroxide (Taliaferro).

The cherts and iron and manganese deposits in the Jurassic Franciscan formation of California have been explained as related to hot springs as sources for the silica, iron, and manganese by Lawson, Davis, and Taliaferro, and Collins and Quirke applied the same explanation to the banded silica formations in the Michipicotin Ranges of Canada, and the latter concluded that the iron and silica were derived from "ascending heated mineralized waters at any loci in a land of great volcanic activity," the silica forming the banded chert formations as deposits in bodies of fresh water and the iron replacing volcanic tuff. The waters are assumed to have contained carbon dioxide, silica, iron, sulphur compounds, and other substances. Not all agree to this explanation (Moore).

Bruce considers that the iron and silica in the North American Pre-Cambrian iron formations were probably derived partly from volcanic rocks, but points out that the several deposits in the Lake Superior Region probably were not all formed the same way. He notes further that the Lake Superior iron formations are not closely associated with volcanism.

Not every one has accepted this explanaton of origin of the Lake Superior formations, and Gruner has urged that the iron in the Biwabic formation is a product of normal sedimentary processes. He postulates that under the conditions of a warm humid climate with an abundance of low vegetation the Keewatin volcanics decayed and released iron oxide, which was readily transported in colloidal form with stabilization by organic colloids. The suspended and tractional noncolloidal sediments were deposited in deltas concurrent with loss in velocity of the transporting waters, but the iron and silica passed onward to clearer waters to be precipitated and deposited largely through the agency of bacteria and algae. Occasional exceptional conditions brought clastic sediments to the sites of deposition of iron and silica, where they may have been subject to rapid deposition, thus becoming interbedded with iron and silica, of which the deposition may have been very slow. Many of the colloidal precipitates assumed oölitic or pisolitic structure, and others became spherulites. Carbonates other than iron were transported still greater distances from the shore, and, if the atmosphere was rich in carbon dioxide, no difficulty of transportation of the iron as the carbonate would have been felt. Rises of water level would have led to migration of the areas of iron deposition over those of clastic sediments, whereas falls of sea level would have produced the opposite effect.

If the Pre-Cambrian and similar iron-bearing deposits are of normal sedimentary origin, it is obvious that very special environmental conditions were required to produce them. But there is nothing strange about this, as special environmental conditions are required to produce the formation of extensive deposits of sandstones—like those that formed the thick Pre-Cambrian quartzites—the evaporites, and probably every other sedimentary deposit. It may well be that the Pre-Cambrian iron formations derived their iron and silica from magmatic waters, but it should not be considered certain that a normal sedimentary origin is precluded.

Very extensive deposits of sedimentary Pre-Cambrian hematites are in the state of Minas Geraes in Brazil (Leith and Harder). These consist of beds of pure hematite and ferruginous schist interbedded with an iron oxide-bearing rock known as itabirite which Tyler has shown is probably recrystallized chert and not quartzite as it has been identified. The iron-bearing formation ranges in thickness from less than 6 to more than 1,200 meters, and some beds of hematite are stated to be 600 meters thick.

These deposits are explained as having been formed through normal processes of sedimentation, with precipitation brought about by normal chemical and biologic processes. As this is one of the greatest and purest deposits of sedimentary iron minerals and as a magmatic origin for the iron seems precluded, it would seem that no great difficulty should be felt in applying the same explanation to other Pre-Cambrian iron formations.

Woolnough suggested that banded iron and silica deposits might be made during times of extreme peneplanation under which condition in a region of dry climate low areas might become covered with salt water to which few carbonates would be brought and banded iron and silica deposits would be formed. Iron and silica under those conditions might be the only sediments contributed by the land, and the banding would arise because of silica being first precipitated and the iron later. This might arise in connection with a wet season succeeded by a dry one.

IRON CARBONATE

Siderite is the only iron carbonate. Outside of isolated and vein forms of siderite, the iron carbonate of particular significance is in the form of concretions, lenses, and beds. Some of the iron carbonate is evidently primary, whereas other occurrences are obviously replacement. Deposition is due to organisms (chiefly bacteria and algae), to removal of carbon dioxide from the transporting medium, and to replacement of other substances. The iron carbonate deposits may be placed in the six classes of concretionary ironstones, blackbands, block ores, limestone ores, oölitic carbonates, and Lake Superior cherty iron carbonates.

The concretionary ironstones, also known as "kidney ores," are spherical to disk-shaped concretions that are composed of more or less iron carbonate together with clay, calcite, and small quantities of other substances. Dimensions range from very small to 30 cm. or more in diameter. The structure is concentric with the laminae in some concretions composed of mixtures of the various substances; in others, some laminae are composed of iron carbonate; and in still others, of different substances. The concretions are found in more or less carbonaceous clays and shales, in which they tend to be arranged in bands, and in some instances a band of concretions may pass laterally into a much cracked layer of clay ironstone.

Blackband iron carbonate derives the name from the color. This is due to contained organic matter, and, like the clay ironstone concretions, the blackbands are interbedded with clays and shales that contain more or less organic matter. Beds are not thick, and the common range is from less than 1 to 10 cm. or more.

Block ores are named from their broken blocky form. Beds are only a few centimeters thick, and they are interstratified with sandy clays and shales. The iron carbonate is impure, and the most common impurity is sand.

Limestone iron carbonate bodies are irregular in form and are thicker units than the blackband and block ores varieties. Occurrences are near the tops of limestone formations, and the ores are thought to have attained position through replacement of limestone. This variety has greater commercial importance than any of the preceding.

The fifth variety is the oölitic iron carbonate and the associated chamoisites in the Cleveland Hills of the Jurassic of England. Some oölitic carbonate is also present in the Brassfield ores (Silurian) of eastern Kentucky. The oölites are up to 1 mm. in diameter.

The sixth variety is the cherty iron carbonate of the Lake Superior Region. This occurs in original form where the iron formations have been protected from oxidation. Colors are light gray, brown, or green, and the texture is very fine grained. Fresh rock consists of alternate laminae of chert and iron carbonate, or argillite and iron carbonate. The origin of these iron carbonates has already been considered.

It is thought that the ironstone concretions, blackbands, and block ores are syngenetic, but they may be penecontemporaneous. The Lake Superior carbonates are syngenetic. The limestone ores and the oölitic ores are probably epigenetic. The iron may have been transported as the bicarbonate and precipitated by bacteria, algae, or other agency as carbonate in an environment deficient in oxygen. If the carbonates are penecontemporaneous, it is possible that the iron was first deposited as ferric oxide and that deoxidation and carbonation followed burial.

IRON SULPHIDES

There are four sulphides of iron that are known to be formed by sedimentary processes: hydrotroilite, melnikovite, marcasite, and pyrite. Melnikovite, marcasite, and pyrite have the same chemical composition, but pyrite and marcasite differ in crystal habit, and melnikovite is probably a colloid.

Hydrotroilite is an amorphous hydrous monosulphide of iron, of which the formula is generally expressed as $FeS \cdot nH_2O$. It may be a hydrophylic colloid. The substance is black, and it forms in black muds and sands containing organic matter and sulphate waters. Formation seems most common in marine waters of poor circulation, as lagoons, seas with weak tides, and inland seas like the Black Sea and Lake Baikal. The blackness of the sediments to a greater or less extent is due to hydrotroilite, and the greater the sulphide content the blacker the color. The mineral may have a considerable geologic significance as it may be a stage in the formation of marcasite and pyrite at one extreme and hydrous and anhydrous ferric oxide at the other.

Melnikovite is the black disulphide. It is not known to be of common occurrence. It has been described as present in streaks, thin layers, and small nodules of Miocene brown clay in the region of Samara, Russia. The nodules have various shapes and concentric structure and rarely exceed 1 mm. in diameter. Some pyrite is associated, and the melnikovite contains quartz, calcite, and flakes of mica.

Sedimentary marcasite and pyrite are present to a greater or less degree in all kinds of sedimentary rocks. Usually the occurrences are in the form of isolated crystals or crystal aggregates in cherts and flints, shales, sandstones, and limestome. Less common occurrences are nodules, lenses, and beds in clays, shales, and coal. Pyrite and marcasite in coal are known as "coal brass." Many fossils in black shales and to a less degree in other rocks are replaced by pyrite or marcasite. The Wabana hematites of Newfoundland contain beds of oölitic pyrite that range in thickness from less than 1 to about 30 cm. These are separated by fissile black shale. The beds of pyrite are persistent but have considerable variation in thickness. Presence of fossils indicates deposition under marine conditions. Similar pyritic strata are reported in the Devonian of Westphalia, where beds are stated to range to 3 meters or more in thickness. It is considered probable that the oölitic pyrite is secondary.

Acording to Newhouse, the iron sulphide layers in coal are chiefly marcasite, whereas those in other rocks are chiefly pyrite.

Concretions are stated to have an inner fine-grained zone that is composed of equidimensional crystals of pyrite and an outer zone of elongated crystals of the same mineral.

Times of origin of melnikovite, marcasite, and pyrite are not definitely known, and probably some are syngenetic and some epigenetic. Replaced fossils are certainly epigenetic. The marcasite and pyrite studied by Newhouse were considered syngenetic with the sediments that enclose them and as having been deposited originally as colloidal gel.

According to Allen, Crenshaw, and Johnston, the character of the medium determines whether melnikovite, marcasite, or pyrite is deposited. Marcasite is said to form in an acid solution, melnikovite in an alkaline one, and pyrite in one that is neutral or slightly acid. The optimum conditions for the formation of marcasite seem to be in a solution containing 1.18 per cent free sulphuric acid at 100°C. Decrease in acidity to neutrality and raising of temperature favor formation of pyrite. Melnikovite seems to form best at temperatures below 100°C., but it may form at 100°C. in alkaline solutions. Environmental conditions that lead to variations in the character of the medium —not rare conditions—would lead to corresponding variations in the sulphides deposited, and a deposit composed of alternate laminae of different iron sulphides might be made.

Bacteria are concerned in the formation of the sulphides of iron. Harder states that bacteria influence the formation of iron sulphide in four distinct ways:

1. Certain bacteria produce decomposition of sulphur-bearing proteins and form hydrogen sulphide.
2. Sulphides of any origin may react with carbon dioxide and water to form hydrogen sulphide.
3. Certain bacteria may act directly on free sulphur to form hydrogen sulphide.
4. Sulphate-reducing bacteria, among which are *Vibrio hydrosulfureus* and *Bacterium hydrosulfureum ponticum* of the Black Sea, in the presence of decaying organic matter take oxygen from sulphates, thiosulphates, and sulphites to form sulphides.

The hydrogen sulphide formed in 1 to 3 may react with ferrous salts in solution to form ferrous sulphide.

Black muds containing organic matter and sulphate waters form the environment of the *sulphuretum* (Fig. 46) of Becking

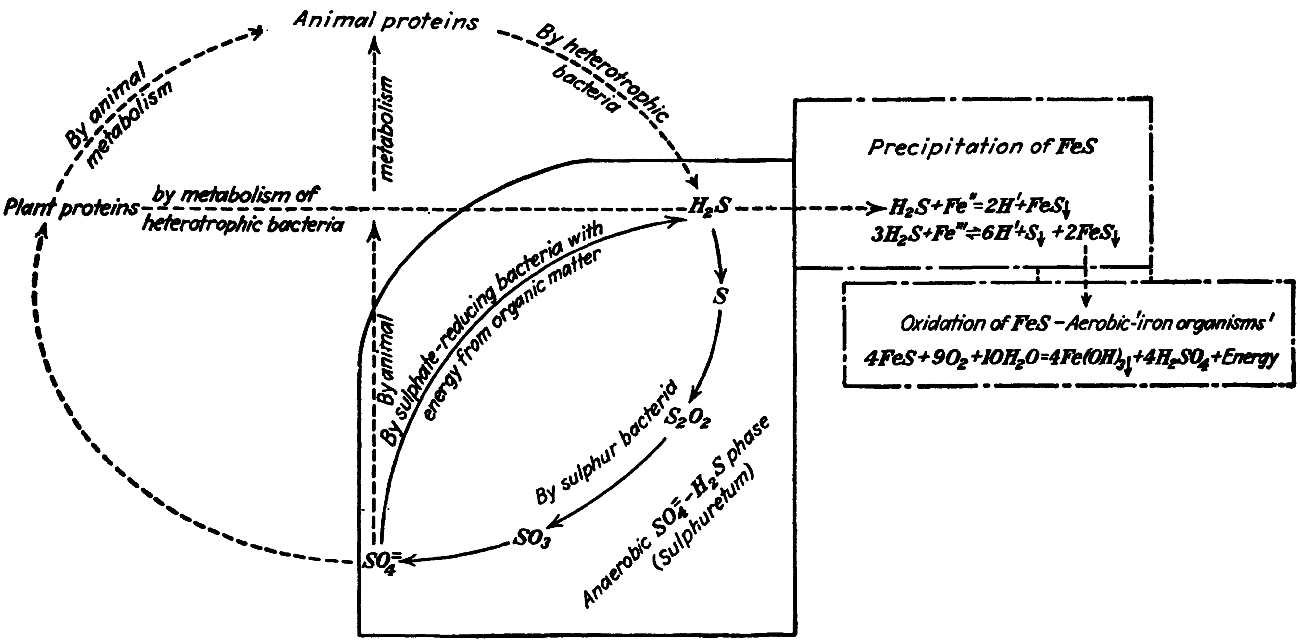
Animal proteins
By heterotrophic bacteria
By animal metabolism
metabolism
Plant proteins
by metabolism of heterotrophic bacteria
H_2S
Precipitation of FeS
$H_2S + Fe'' = 2H' + FeS\downarrow$
$3H_2S + Fe''' \rightleftharpoons 6H' + S\downarrow + 2FeS\downarrow$
Oxidation of FeS - Aerobic 'iron organisms'
$4FeS + 9O_2 + 10H_2O = 4Fe(OH)_3\downarrow + 4H_2SO_4 + Energy$
By sulphate-reducing bacteria with energy from organic matter
By animal
S
S_2O_2
By sulphur bacteria
SO_3
$SO_4^=$
Anaerobic $SO_4^=$-H_2S phase (Sulphuretum)

FIG. 46. Diagram illustrating the sulphur cycle in sediments, with special consideration of the reactions concerned in the formation of ferrous sulphide. Reactions are also given leading to the formation of ferric from ferrous iron. The ferric iron forms a brown layer overlying the sulphide-bearing sediments. [*After E. W. Galliher, The sulfur cycle in sediments, J. Sed. Pet.*, 3 (1933), 52.]

and Galliher. This is anaerobic bacterially and alkaline chemically except at the top of the muds where aerobic conditions may prevail and ferric hydroxide be formed. It is in the sulphuretum that the sulphate-hydrogen sulphide-sulphate cycle takes place. According to Galliher (1933), at least five different kinds of bacteria live in the sulphuretum. These are: (*a*) sulphide bacteria, which obtain their energy from oxidation of hydrogen and other sulphides, (*b*) thiosulphate or thionic acid bacteria, which obtain energy from oxidation of thiosulphates, (*c*) sulphur bacteria, which obtain energy from oxidation of sulphur, (*d*) two associated organisms of which one oxidizes sulphur to hyposulphite (S_2O_2) and the other oxidizes the hyposulphite to the sulphate (SO_2), and (*e*) bacteria that reduce sulphates and other oxygen-rich compounds, as thiosulphates, to hydrogen sulphide.

Surface of the water

↓ ↓ ↓
Oxygen diffusing from air into water
↓ ↓ ↓
Level of optimum conditions for hydrogen sulphide–consuming bacteria and formation of sulphur and sulphuric acid
↓↑
↑↓Sulphur and sulphate settling to the bottom and↓↑
hydrogen sulphide rising
↓↑
Vicinity of the bottom and in the bottom deposits with sulphate-reducing and hydrogen sulphide–forming bacteria

---

FIG. 47.

Group *d* bacteria acquire energy from carbohydrates or amino acids, and the rate at which they work depends upon the nature of the source of the carbon. The bacteria of group *e* are strictly anaerobic, deposit sulphur in their tissues, and require essential inorganic matter in available forms. Groups *a*, *b*, and *c* are autotrophic. They do not require organic matter and use energy liberated by inorganic reactions. Groups *d* and *e* are heterotrophic and obtain energy from carbonaceous sources. Galliher gives a diagram (Fig. 46) that shows the reactions that take place in the sulphuretum to form the ferrous sulphide and in the surficial brown layer to form ferric hydroxide. The sulphur cycle from reduction of sulphates to formation of hydrogen sulphide and sulphuric acid and reformation of sulphates is shown in Fig. 47.

SEDIMENTARY IRON SILICATES

The common sedimentary iron silicates are greenalite, glauconite, celadonite, chamoisite, and berthierine. Greenalite and glauconite have extensive geological distribution; the others seem to be relatively rare.

Greenalite. Greenalite, except for occurrences in the Ordovician of Scotland (Kennedy) and Liassic sandstones of Sweden (Hadding, 1929), is known only from the Pre-Cambrian of the Lake Superior Region, where it is considered by many geologists as the original substance from which some of the Lake Superior hematites were derived. It is generally assumed to have been formed from reactions of alkaline silicates and ferrous salts. The composition is stated by Joliffe to be $3FeO \cdot 4SiO_2 \cdot 2H_2O$, and it is in the form of amorphous dark-green spherules that range in shape from spherical to ellipsoidal and in dimension to about 1 mm. There is no cleavage or crystal form. The spherules are commonly embedded in a matrix of chert.

Mead formed substances resembling greenalite through reaction of ferrous sulphate or ferrous chloride and sodium silicate in saline solution (Van Hise and Leith). The precipitates had green color, granular texture, and the optical properties of greenalite from rocks of the Mesabi Region of Minnesota. Equations expressing the reactions are as follows:

$$FeSO_4 + Na_2O \cdot 3SiO_2 = FeO \cdot 3SiO_2 + Na_2SO_4$$
$$FeCl_2 + Na_2SiO_3 = FeSiO_3 + 2NaCl$$

In nature the sodium silicate might be derived from the weathering of igneous rocks and from contact of lavas or emanations therefrom with sea water. Studies by Joliffe suggest that the constituents of greenalite are of magmatic derivation.

Glauconite. Glauconite is the most common sedimentary iron silicate, and the mineral is present in sediments from the Cambrian to those now being deposited over many parts of the ocean floor. It is usually an accessory substance in sediments, and these consist of all kinds except evaporites. In a few cases it forms the dominant constituent of a bed. It is most common in sandstones and limestones, particularly the former, whence the appellation, "greensand."

Glauconite is not certainly known to form in other than marine waters, and its presence in sediments is generally considered good evidence that the waters of deposition were marine. Celadonite, which forms as a decomposition product, has somewhat the same appearance and was considered by von Gümbel and Glinke as chemically identical with glauconite, but Hendricks and Ross have shown that it is a distinct substance, commonly forms in vesicular basalts under reducing conditions, and derives its magnesium, iron, and silicon dioxide from olivine and its other constituents from deuteric water. Glauconite, on the other hand, is formed only in a marine environment under reducing conditions maintained by bacterial action. It derives its magnesium and potassium from sea water and the other constituents from muds. Both are heptophyllic micas.

Glauconite is present in sediments as irregularly spheroidal particles with botryoidal surfaces and minute particles ranging to microscopic dimensions. There is no definite internal structure, and nothing resembling concentric lamination has been observed (Fig. 48).

The composition of pure glauconite is said to be $FeK \cdot Si_2O_6 \cdot nH_2O$. The iron may be partly replaced by alumina, and some of the potassium may be replaced by other bases. Schneider expressed the composition as $(KNaFe''Mg) \cdot (Fe'''Al) \cdot Si_6O_{18} \cdot 3H_2O$. This shows ferrous as well as ferric iron and a definite quantity of water. Ross states that glauconite represents a series having the form of $R_2O:RO::R_2O_3:SiO_2$, which at one extreme has the substances in the proportion of 1:1::2:10 and, at the other 1:1::3:10. Analyses of glauconite are given in Table 24. An analysis of celadonite is also given. Hendricks and Ross give representative formulas of glauconite and celadonite.

Glauconite rarely, if ever, makes a pure deposit. Associated substances are quartz, white mica, clay minerals, feldspar, hornblende, magnetite, garnet, epidote, and various shell and rock fragments. Usually the various other substances dominate, but there are occasional thin layers that are largely glauconite. Many occurrences contain small concretions of calcium phos-

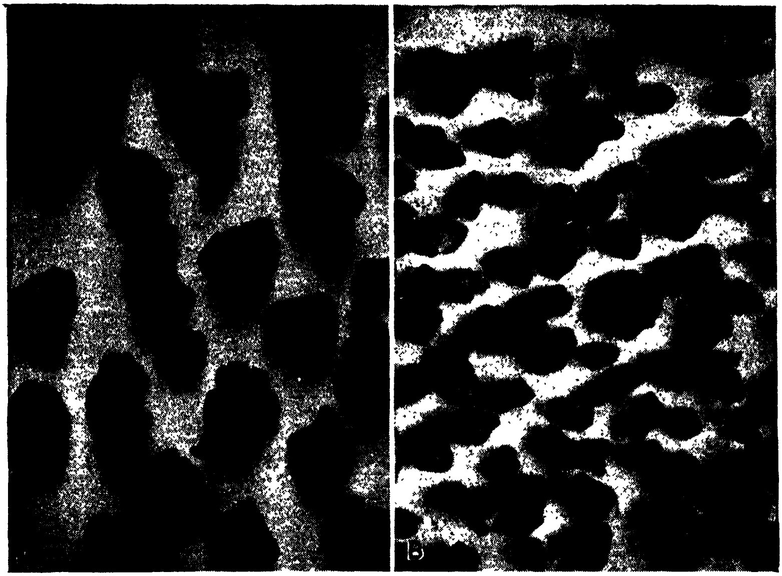

FIG. 48. Microphotographs of glauconite from the Cretaceous of New Jersey. *A*, grains ranging in diameter from 1¼ to 2½ mm. in diameter. These are maximum dimensions. *B*, grains showing considerable wear, dimensions range from ⅝ to 1¼ mm. In both cases magnified 9 times. (*Photograph by G. R. Mansfield, U.S. Geol. Surv.*)

TABLE 24. ANALYSES OF GLAUCONITE AND CELADONITE

Substance	Per cent[1]	Per cent[2]	Per cent[3]
SiO_2	53.61	54.84	48.50
Al_2O_3	9.56	3.52	9.00
Fe_2O_3	21.46	12.64	20.00
FeO	1.58	4.90	3.10
MnO	trace	0.24	
MgO	2.87	6.65	3.70
CaO	1.39	0.89	0.40
K_2O	3.49	7.00	6.20
Na_2O	0.42	0.39	1.50
H_2O	5.96	9.62	6.40
Total	100.34	100.69	98.80

[1] Mean of four analyses of glauconite collected by the "Challenger" expedition.

[2] Mean of four analyses of celadonite from Scotland.

[3] Franconia formation of the Upper Cambrian of Wisconsin, H. Schneider, *J. Geol.*, 35 (1927), 296.

phate. Calcite seems invariably present in the form of shell fragments. It is the presence of these various substances within and adhering to glauconite particles that makes analyses of uncertain value.

The association of glauconite with other substances is obviously mostly one of secondary deposition, as the deposits are cross-laminated and the glauconite particles show abrasion. Such is evidently the case for most of the occurrences in the Upper Cambrian of the upper Mississippi Valley.

Collet states that in existing marine waters of the sea glauconite has a depth range from 82 meters on the west coast of Africa to 2,512 meters in the Indian Ocean. According to Galliher it begins to form in Monterey Bay, California, in depths of 5 or 10 fathoms and becomes abundant at the depth of 50 fathoms. It is rare in the deep ocean basins, and in the present seas it seems that rich glauconite occurrences lie adjacent to land areas where plutonic and metamorphic rocks are exposed.

There has been much speculation respecting the origin of glauconite. For a long time it was thought that foraminiferal and other small shells were essential for formation. It was assumed that shells became filled with mud and organic matter and that decay of the latter led to transformation of the iron in the mud to the sulphide, with ultimate formation of sulphuric acid. The acid decomposed the clay to colloidal silica and alumina. The latter was removed. Ferric hydroxide formed from the iron sulphide, and colloidal silica absorbed potassium from the surrounding sea water. Glauconite resulted.

Collet suggested that hydrogen sulphide united with iron oxide to form ferrous sulphide, which reacted with colloidal matter in clay. Water and potassium were then absorbed from sea water. Three stages were postulated: The first stage had small shells filled with clay. In the second the iron in the clay oxidized, the alumina was expelled, and its place was taken by ferric oxide. Potassium and water were absorbed in the third stage. It was postulated that the absorption of potassium continued indefinitely and that older glauconites were richer in potassium than younger. This is not supported by analyses.

Each of the older hypotheses relates glauconite formation to shells of foraminifera, but examination of glauconite by many students gives little support to this view although Dryden noted that glauconite in the Miocene Calvert formation of the Atlantic Coastal Plain is found only in foraminiferal shells and that glauconite in such shells is also present in the Vicksburg of the Mississippi Region and the Eocene of the Paris Basin. This, however, does not prove a causal relation. Quite recently it has been suggested that there is a relation between glauconite and excremental particles. Takahashi and Yagi collected pellets in fresh waters closely connected with the sea that are said to show all transitions from those without a trace of glauconitization to those completely glauconitized.

Galliher (1935) reverses the process of formation from that of the building up of a mineral to that of the end product of the destruction of one. According to Galliher, the initial stage leading to the formation of glauconite is a particle of biotite. This undergoes decomposition on its exterior and along cleavage planes. Swelling takes place, and the particle loses its micaceous form to become a more or less rounded prism which on further decomposition changes to various shapes. Complete alteration leads to disappearance of the biotite and to the amorphous internal structure characteristic of glauconite.

Galliher found that the alteration began to take place at depths of 5 to 10 fathoms, that the quantity of biotite decreased and that of glauconite increased with depth, and that at a depth of about 50 fathoms the biotite had nearly disappeared and alteration was essentially complete. There seemed to be an intimate association with areas of little agitation, bottoms covered by black muds, and anaerobic conditions. Oxidation of the iron in the biotite to the trivalent state was assumed to be as expressed in the following equation:

$2FeO + H_2O = 2H + Fe_2O_3$, with fermentation processes, nitrate reduction, and sulphate reduction functioning as hydrogen acceptors. The environment is alkaline, and the alkalinity may reach a pH of 9 or 10.

The time factor is probably very important in the formation of glauconite, and, as emphasized by Hadding, glauconite does not tend to form except under conditions of slow or negative deposition. Under conditions of rapid deposition, as long ago pointed out by Murray and Renard, glauconite either does not form or forms only to a slight extent. The parent mineral, assumed by

Galliher to be biotite, must be exposed for a long time to the diagenetic processes.

Galliher summarizes his views as follows: The change of biotite to glauconite involves loss of some alumina, potassium, and magnesium; oxidation of iron; and a gain of water. The change is favored by marine conditions in an alkaline solution, an anaerobic environment, and slow deposition of sediments, so that diagenetic processes may continue for a long time. The absence of glauconite from deep-sea deposits is explained as due to the inability of biotite to reach the deep sea because of deposition near shore.

Examination of glauconite from horizons ranging from Cambrian to Recent and from many parts of the world led Galliher to generalize that in all cases studied the genesis of glauconite is tied up with alteration of biotite and that the relation to foraminiferal shells is incidental and not causal.

As pointed out by Allen and others, there are some occurrences of glauconite that are difficult to explain as due to biotite alteration, from simple inability of biotite to enter the waters of formation, or be present in the quantity required to form the vast quantities of glauconite that exist in parts of the geologic column. Allen's views relate to the Glenwood member of the Platteville formation of the upper Mississippi Valley and the Bonneterre formation of Missouri. No metamorphic minerals are present in these strata; there is great scarcity of heavy minerals; and location of source rocks to yield the biotite is difficult. The relations in the Upper Cambrian of the upper Mississippi Valley are also difficult to explain on the hypothesis of alteration of biotite. There is nothing in these strata indicating anaerobic conditions. The conditions were evidently those of aeration, yet the Franconia formation contains an abundance of glauconite, and it is common in the Eau Claire member of the Dresbach formation. The Trempeleau formation of essentially the same depth of water and deriving its sediments from the same source contains little indigenous glauconite. The New Jersey deposits of glauconite do not seem possible of explanation on the hypothesis of Galliher. It seems doubtful if his explanation has general application although it may apply to the glauconite now forming in Monterey Bay.

Chamoisite. Little is known of the origin of chamoisite. It consists of about 14 per cent SiO_2, 7 to 8 per cent Al_2O_3, around 60 per cent FeO, and about 18 per cent H_2O. The mineral is present in compact or oölitic form and has a greenish-gray to black color. Magnesium is present in some. Berthierine has similar occurrences and has color of greenish black to bluish gray. The composition is much the same as that of chamoisite, but there are about 75 per cent FeO and about 5 per cent H_2O.

The two minerals are more or less associated. Chamoisite is present in the Wabana ores of Bell Island, Newfoundland, in the Jurassic of the Cleveland Hills, England, the Minette ores of Lorraine in France, and elsewhere. Berthierine is found in the Minette region of France. Both are placed in the chlorite group by Dana, and it seems likely that the two substances are products of diagenesis.

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SEDIMENTARY MANGANESE

INTRODUCTION

Deposits of manganese minerals are not common, although the element is a minor constituent of most sedimentary rocks. The element has at least five valences and behaves very much like iron in transportation and methods of deposition. The many valences make numerous compounds possible of which some are soluble. Manganous compounds are more soluble than those in which manganese has higher valence.

The original source of manganese in sediments is the igneous rocks, in which it is present in very small quantities. It is released from this source on weathering and probably takes some form of manganese oxide. It may be transported as the bicarbonate in water with a high content of carbon dioxide, and such may have been the common method of transportation in Pre-Cambrian periods if the earth's atmosphere was then at any time largely composed of this gas and had a low oxygen content. Under existing conditions it is probable that the prevalent method of transportation is as manganese oxide hydrosols stabilized by organic colloids. What has been said of transportation of iron and silica is applicable to some extent to manganese. The quantity in stream waters is very small, commonly ranging from less than 0.5 to about 5 parts per million in the river waters of the Temperate Zone, although 117 parts per million have been noted in spring waters. The quantity in the sea is extremely small, but it has been detected. Evidently it behaves like iron and silica and is largely precipitated shortly after reaching the sea. As manganese seems to be required in the functioning of organic matter, it is present in most, if not all, organisms and thus has an organic cycle not unlike that of iron, phosphorus, and some other elements, in that much is in continued organic circulation from which escape is possible only under particular conditions.

PRECIPITATION OF MANGANESE

It seems probable that deposits of manganese are formed in ways similar to those of iron. If manganese is carried as the bicarbonate, liberation of carbon dioxide leads to precipitation as the carbonate. It is possible that bacteria may take carbon dioxide from the bicarbonate and precipitate the carbonate, and carbon dioxide may be taken from manganese carbonate through oxidation and the manganese precipitated as the oxide. Manganese transported as the oxide hydrosol may have the protective colloid removed by bacteria and the oxide precipitated, or the precipitation may take place through electrolytic action just as for iron, but the manganese oxide is precipitated after the iron, as laboratory experiments indicate that under several conditions manganese oxide precipitation is slower than ferric oxide precipitation.

Experimental work has demonstrated that several genera of bacteria (Baier, Thiel), common in oceanic muds, particularly *Crenothrix*, *Leptothrix*, *Cladothrix*, and *Clanothrix*, precipitate manganese oxide from a number of manganese salts, and water mains connected with city water supply are known to have become clogged with impure manganese oxide considered to have been precipitated by bacteria. Vernadsky expressed doubt that oxygen alone at ordinary temperatures could change manganous compounds to higher manganese oxides, and he believed that bacteria were required to bring about the change. This extreme position ignores laboratory work to the contrary. Algae have been considered as agents of precipitation of manganese oxide, and it has been suggested that concretionary forms of manganese oxide, such as those found in beds of manganese oxide near Tschiaturi, Russia, may have been deposited by algae. It has been shown that certain algae extract manganese oxide from the waters of the Elba River, of which the manganese content is only 0.25 to 0.65 part per million, and algae have been used as a basis of processes to remove manganese from waters. David White has stated that during times of drought algae in Furnace Creek, W. Va., seem to deposit manganese oxide. Fungi are also said to precipitate manganese compounds.

MANGANESE DEPOSITS

Sedimentary deposits of manganese compounds are placed in the two classes of primary and secondary, but many deposits seem gradational. The deposits are mainly carbonates and oxides, and the common minerals are manganite ($Mn_2O_3 \cdot H_2O$), pyrolusite (MnO_2), psilomelane (H_4MnO_5), and rhodochrosite ($MnCO_3$). Braunite ($3Mn_2O_3 \cdot MnSiO_3$) and hausmannite (Mn_3O_4) are more rare. Manganese oxides formed in bogs and lakes are commonly known as *wad*.

Primary Deposits of Manganese

MANGANESE CARBONATES

Thin zones of manganese carbonate and manganese carbonate concretions seem to be fairly common in fine-grained marine sediments. Such are the thin layers of green jaspery carbonate in the Cambrian (Dale) near Trinity and Conception bays, Newfoundland, in the midst of thinly bedded red and green shales. In the Manuel River section there are four persistent and several lenticular layers containing manganese carbonate. The range in thickness is from 6 to 21 cm. The section at Brigus has six beds containing manganese carbonate with a total thickness of 1.35 meters. Manganese carbonate composes from 10.23 to 44.30 per cent of the beds, and manganese oxide ranges from 2.34 to 28.93 per cent. It is not known whether the oxide is primary or secondary. A layer exposed on Placentia Bay contains 84.6 per cent manganese carbonate. The Cambrian of Merionethshire, Wales, and near Chevron, Belgium, contains similar layers of manganese carbonate, and one layer at the Belgium locality is said to be about 1 meter thick.

The Franciscan chert of California contains thinly laminated beds of gray, nearly pure manganese carbonate with thickness ranging from 60 to 180 cm. Laminae in the beds range from 2 to 5 mm. thick, and the manganese carbonate is in the form of minute spherules cemented together. Occasional organic remains show a marine origin. If the interpretation of Lawson, Davis, and others that the silica forming the cherts is of magmatic origin is correct, it may be that the manganese had the same derivation. The Ordovician Cason shale of the Batesville district of Arkansas contains nodules of iron-manganese carbonate. These resemble concretions, and Ulrich considered them algal structures of the genus *Girvanella.* Unoxidized nodules have the manganese entirely in the carbonate form. They contain from 12 to 20 per cent manganese and 7 to 10 per cent iron. The Pre-Cambrian of the Cuyuna Range of Minnesota contains cherty manganese-bearing iron carbonate with as much as 30 per cent iron and 5 per cent manganese. A magmatic source has been postulated for the iron and silica.

The Pierre shales (Cretaceous) near Chamberlain, South Dakota (Zapffe), have a 10-meter zone that contains several persistent layers of manganese-iron carbonate concretions or nodules. The concretions are lenticular or flattened spheroidal, 7.5 to 20 cm. in diameter, commonly have organic nuclei, and contain about 16 per cent manganese and 10 per cent iron. The manganese content is twice as great in the center as in the peripheral zone. On exposure the iron oxidizes in advance of the manganese.

Deposits of manganese in residual deposits of eastern United States, derived from weathering of definite strata in a wide range of Paleozoic rocks, and similar deposits of manganese oxide in the underlying rocks to depths of as much as 120 meters beneath the surface are interpreted as having been produced by oxidation either of beds of rather pure manganese carbonate or concretionary zones of iron-manganese carbonate with little or no oxide in either. There is no direct evidence for this conclusion other than the rocks from which the manganese is thought to have been derived resemble rocks that contain primary manganese carbonates and do not resemble those that contain primary oxides of manganese.

MANGANESE OXIDES

Manganese oxide is now in process of deposition in the sea, where it is present as finely divided coloring matter in some deep-sea sediments, coatings over objects on the bottom, and as nodules and concretions. Some manganese oxide is present in nearly every sample of deep-sea sediments, and over some shallow areas the sediments are colored reddish at the surface from iron and manganese oxides. The red clays of the Pacific Ocean are darker than those of the Indian and these in turn are darker than those of the Atlantic by reason of the higher content of manganese oxide. This perhaps is correlatable with the extent of volcanic contributions to the three bodies of water.

Coatings of black manganese oxide may be found on shells, bones, teeth, pebbles, and boulders, and these may be present in both deep and shallow water. Boulders in places on the shores of Change Island in Notre Dame Bay, Newfoundland, are coated with a black shiny layer of manganese oxide that has the appearance of desert varnish.

Nodules of manganese oxide (mostly MnO_2) are present in all depths on some parts of the ocean bottom, and they are also present on the bottoms of some fresh-water lakes. The nodules range in shape from spherical to irregular and in dimension from very small to 15 or more cm. in diameter. The surfaces tend to be mammillary, and shapes to some extent depend upon shapes of nuclei, which are commonly fragments of pumice or other volcanic rock more or less replaced by manganese oxide. Analyses show variable but important percentages of silica, alumina, ferric oxide, and calcium carbonate with a range in manganese oxide from 25 to 35 per cent and manganese carbonate from about 1.5 to 3.5 per cent. The manganese oxide is often equaled and in some cases exceeded by the iron oxide.

Parts of the bottom of Trout Lake, northern Wisconsin, to a known depth of about 18.5 meters contains nodules and in places crusts of a black deposit with a known thickness of 15 cm. The deposits consist of manganese and iron oxides and organic matter, and there is about twice as much manganese as iron.

Manganese nodules of disk and pear shapes, about 5 mm. thick, and as much as 5 cm. in diameter have been dredged from the bottom of Ship Harbor Lake, Nova Scotia, in depths of 0.6 to 1.8 meters. The disk-shaped nodules have a structure concentric to the center and a nucleus that is either a rock or a fragment of a manganese nodule. The manganese oxide content (MnO_2) is 36.08 per cent. The nodules were described by Kindle and ascribed to an algal origin. It seems improbable that an inorganic origin can be postulated (Fig. 16).

Manganese nodules have been collected in Loch Tyne in Scotland and in the Zeller Sea in Austria, where they have been found with diameters as great as 20 cm. Depths of derivation were as great as 20 meters.

Beds of pure manganese oxide have been found in Russia, Cuba, Nevada, India, Brazil, and elsewhere. None is more important than that of the Tschiaturi District of Russia, where a single bed in Tertiary strata extends over an area of 130 square kilometers, and there are also other beds of large areal extent. The manganese oxides of this district have oölitic texture and lie in the midst of arkosic sediments. The manganese bed of greatest extent varies in thickness from 1.8 to about 5 meters, and the manganese oxide lenses range in thickness from 7.5 to 60 cm. The oölites are flat ellipsoids with larger diameters from 1 to 10 mm. and the average about 3 mm. Nuclei of quartz or feldspar are present in most. Composition seems to be that of pyrolusite or psilomelane. The oölites are embedded in arkosic materials with which they were deposited, and as they are rudely sorted and contain little or nothing of the materials with which they are associated, it seems that they were formed elsewhere and attained present positions after transportation. Most oölites seem to have shrunken from the matrix and thus loss of water is indicated. Places of origin are not known.

Some manganese oxide deposits of Cuba resemble those of the Tschiaturi District. They are associated with Tertiary sediments, and some contain spheroidal particles of manganese oxide as much as 6 mm. in diameter. Beds to 6 meters thick have a manganese content ranging from 10 to 20 per cent. The Charco Redonda mine has a 45-cm. bed composed of nearly pure manganese oxide in the form of small cauliflower growths that resemble those made by algae.

Pliocene latite tuffs near Las Vegas, Nev., contain beds of wad with thickness as great as 11 meters. Beds of nearly pure wad are also present in lake clays and fine sands near Cleveland, Idaho.

Secondary Manganese Deposits

Pyrolusite and psilomelane and some other oxides of manganese are very stable in the zone of weathering under most conditions, but most other manganese minerals are soluble, so that sulphides, silicates, carbonates, and phosphates of manganese are quickly oxidized, dissolved, and redeposited as flocculent hydrous oxide similar to wad, and, with continued deposition, psilomelane, manganite, and pyrolusite tend to form in the order given. The surface zone contains mostly coherent psilomelane, and the deepest zone of secondary deposition contains only earthy wad. As the higher manganese oxides are quite insoluble and very stable, they tend to accumulate in the residuum, and thus deposits of manganese oxide may be found over deeply eroded areas. The manganese may not be concentrated at the places where the parent rocks were eroded.

Environments of Deposition of Manganese

Manganese oxides and carbonates are known to have been deposited under both marine and fresh-water conditions. They seem to be associated to some degree in the geologic column with chert, iron compounds, glauconite, and calcium phosphate, and it is probable that similar conditions are necessary for their formation. However, there are manganese deposits that bear no relation to chert, iron compounds, glauconite, or calcium phosphate. It has been suggested that manganese deposition may bear some relation to unconformities, as some deposits are near the base of strata overlying unconformities. Many, however, bear no such relation, and most unconformities are not immediately overlain by formations that contain important quantities of manganese compounds. Some of the relations to unconformities may be explained as due to concentration of residual materials and others to concentration on the base level of deposition or some profile of equilibrium. Manganese deposits do not commonly occur in strata that have an abundance of fossils, but before this may be given significance it should be demonstrated whether this means there was a paucity of life on the bottom. Manganese carbonate is known in bog deposits, but it is not known whether deposition was first as the oxide and there was later reduction by decaying organic matter to the carbonate. It does not seem to have been deposited in the midst of coarse sediments or in thick limestones. Distribution is from the Pre-Cambrian to the present. Manganese oxides occur in the midst of limestones, marls, tuffs, and shales, in bog and lake deposits, and about hot springs. Distribution is also from Pre-Cambrian to present. The environmental conditions leading to extraordinary local concentration of manganese compounds are unknown.

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CHAPTER XI

SEDIMENTS OF CHEMICAL DEPOSITION

THE CARBONACEOUS SEDIMENTS

INTRODUCTION

Sediments defined as carbonaceous include original organic tissues and subsequently produced derivatives of which the composition is chemically organic. The organic matter may be carbonized (coalified) or bituminized. Carbonaceous sediments are characterized by relative abundance of carbon, which is either free (fixed) or united with hydrogen and oxygen. The sediments are typically carbonaceous if there is a dominance of cellulosic and ligneous materials, and bituminous if rich in substances that produce hydrocarbons. The former are termed humic. There are all gradations between the two extremes.

Carbonaceous sediments are found more or less rarely to abundantly in all kinds of sedimentary deposits since the beginning of the Huronian and they are also present in Pre-Huronian formations. Materials easily recognized consist of stems, leaves, seed coats, fruits, algal thalli, etc., but much is with difficulty, if at all, determinable and consists of minute fragments derived from many kinds of organic materials and derivative products in which no organic structures are apparent. The derivative products are generally designated humic, ulmic, or ulmohumic decomposition products. Most of these are composed of particles of colloidal dimension and may be seen in liquid form in the brown to black waters of peats and in highly diffused form in swamp waters. These finely divided substances ultimately become the "paste" of coals.

Carbonaceous sediments may be placed in the two classes of primary and secondary. The former includes the coals and coaly and bituminous materials distributed in many different kinds of sediments. The secondary products are formed from the first by processes operating subsequently to deposition and consist of solid, liquid, and gaseous hydrocarbons. These may be widely disseminated through the sediments or be locally segregated.

Carbonaceous sediments are produced and deposited on land and in both fresh and salt waters. Those grown and deposited under distinctly aquatic conditions in both fresh and salt waters tend to be of a bituminous character, whereas those grown on the land are mainly coaly. The carbonaceous sediments of streams and many small ponds present transitions between the two extremes.

Carbonaceous materials generally give dark colors to the sediments in which they are present, and the depth of color is a rough measure of abundance. Some bituminous materials are not dark-colored.

The range of carbonaceous materials in sedimentary rocks is from nothing to nearly 100 per cent. The purest deposits are found in coals. Fine-grained sediments generally contain more than coarse-grained. The quantity in calcareous sediments ranges from little to much.

All living organisms constitute the sources of carbonaceous sediments. Plants seem to have been the greater contributors to that which is identifiable, but both plants and animals may be considered to have contributed to the decomposition derivatives.

Carbonaceous materials which grow at the places where the sediments are deposited are *autochthonous;* those transported to the places of deposition are *allochthonous.* The latter are considered of little importance so far as coal deposits are concerned.

FACTORS CONTROLLING PRESERVATION OF ORGANIC SEDIMENTS

Under conditions of exposure to the atmosphere, dead organic matter is eaten by scavengers or undergoes decomposition through the activities of microorganisms. Nothing remains except the hard parts and inorganic materials dissolved in the body fluids and in the tissues. Decomposition is mainly due to molds and bacteria. These microorganisms are probably adjusted to every form of endemic organic debris, and under ordinary conditions no organic matter escapes destruction. If free oxygen is not readily available, decomposition is retarded, but not suppressed, as it may continue to be carried on by molds and anaerobic bacteria. This destruction of organic matter by microorganisms is designated *microbian* or *biochemical*.

In order for organic substances to be preserved it is essential that microbian activity be suppressed. Suppression can be accomplished only by producing conditions destructive to organisms. This may be done by deposition of a cover of clays and silts; by a sufficiently rapid rate of deposition of organic materials to produce rapid burial; perhaps most effectively by accumulation in the sediments and overlying stagnant waters of toxic products made by the decompositional microorganisms, which products in course of time destroy their makers; and by deposition in extremely cold waters. The last may be not uncommon in lakes of high latitudes of which the bottom waters may rarely become warm.

The materials surviving decomposition at the time of suppression of the microorganic activity are the more resistant and decompositional products resulting from microbian activity. Suppression of microbian activity at the time of deposition of organic materials makes possible more or less complete preservation. Retardation merely prolongs decomposition and produces products related to retardation, as methane, ammonia, and hydrogen sulphide. The characters of the carbonaceous sediments ultimately buried are dependent first, upon the quality and quantity produced and deposited, and second, upon the stage attained in decomposition when microbian activity is suppressed. Quiet waters favor early suppression of microbian activity. Much agitated waters with strong currents favor microbian activity and may lead to total disappearance of organic materials. Preservation is not possible where burial is precluded.

Microbian processes are selective and result in early elimination of the less resistant parts of organic materials and concentration of the more resistant parts. First thought to disappear are the proteins, carbohydrates, and little-resistant forms of cellulose, followed later by the lignins and lastly by the fats, waxes, gums, and resins. An organic sediment that originally was low in fatty, waxy, resinous, and gummy constituents may be made rich in these substances through microbian elimination of other constituents.

After biochemical decomposition has been arrested by burial of the organic matter, change is continued by geochemical processes and coalification advances from rank to rank with anthracite, perhaps graphite, as the ultimate goal. Schopf has divided the entire range of the coalification processes into *incorporation*, *vitrainization*, and *fusinization*. According to Schopf, incorporation is that phase of coalification in which the original materials were phyterals of pollen and spore coverings, plant cuticles, fatty algae, etc. The process is postulated as one of slow change. Vitrainization takes place gradually on materials which are chiefly phyterals of lignic and cellulosic character, in contrast to fusinization which takes place rapidly on the same materials.

The Vegetable Matter of Coals

Many coals show an abundance of large plant remains, particularly along bedding planes, in the form of wood, stems, leaves, bark, and roots. In some of the less altered coals the plant remains are so obvious as to lead to the designation of "woody." Under high magnification, wound resin, spores, spore cases, and small fragments showing plant tissue and cells become visible in coals that have not been too highly altered. Wound resin is common and frequently abundant in Mesozoic and Cenozoic coals but is generally not common in the more altered Paleozoic coals. It is probable, however, that Paleozoic coal plants contained as much resinous matter as plants of later time, but that alteration led to elimination. Etching of polished surfaces reveals additional evidence of the contribution of plants to coals and proves that coals are composed of microscopic to large fragments of plants immersed in a "paste" that originally was a derivative of decomposition (Fig. 49).

Concretions, known as "coal balls" or "bullions," are found in some coals (Stopes and Watson). These are composed of carbonate or silica more or less stained by the brown humic derivatives that permeated the peaty deposits in which the "coal balls" formed. These concretions often have nuclei in the form of excellently preserved plant structures.

Shaly or "bony" layers of coal often contain plant remains in extremely excellent preservation. Fusain layers, the so-called mineral charcoal, may also show excellently preserved plant structure.

As the water covers of the coal swamps changed in depth either by drainage, raising of the surface of the bottom by deposition, or because of periods of dry and wet years, there can be little doubt that modification of the plant associations changed also, and evolution of plants also led to changes from period to period. The result is that the characters of plants of any coal bed may change from level to level and over wider limits from the coal of one geologic period to that of another. However, the plant remains of coal indicate that the range of plant life and chemical character of the plant materials throughout much and probably all geologic time were not unlike those of existing plants.

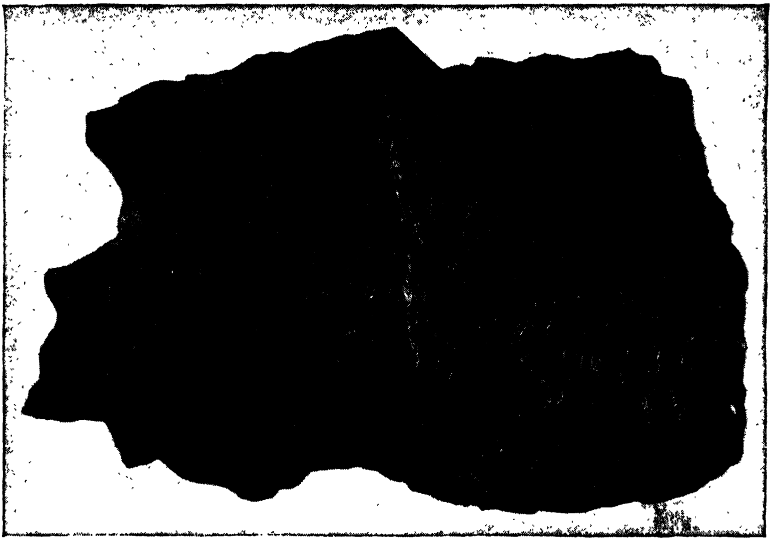

FIG. 49. Photograph, one-half natural size, of the bedding plane of a laminated bituminous coal from Indiana. Miscellaneous organic debris, consisting of fragments of wood, ferns, and other organic materials, cover each bedding plane.

Humic Derivatives Formed in Organic Sediments by Microbian Activity

Microbian activity results in the formation of finely divided and other organic materials that are designated humic derivatives. Under conditions of stagnant water, humic derivatives become concentrated and produce toxicity in the waters. This ultimately arrests and finally suppresses microbian activity and decomposition. It may also be arrested by deposition of lush supplies of organic matter, as such matter excludes air and aerating waters, conserves the toxic products, and thus preserves the organic materials. Microbian activity is favored by warm waters and retarded or suppressed by cold. It is also favored by agitation of the water cover, as this increases the oxygen content, and by much inflow and outflow due to great precipitation or other cause, as humic derivatives and toxicity are removed in this way. As there are some half-dozen variables bearing upon retardation, hastening, or suppression of microbian activity; concentration of toxic products and humic derivatives; and preservation of organic sediments; it is obvious that there is likely to be considerable variation in the character of the organic sediments finally entombed.

The depth to which microbian activity extends in organic deposits is not known. It evidently depends upon the depth to which oxygen penetrates and to which toxic products are developed and retained. In the deposits of lakes there is a rapid falling off in numbers of bacteria a short distance below the surface of the bottom, but spores of bacteria in the resting stage and presumably capable of growing have been reported in a peat deposit to the depth of 9 meters. Below the depth to which microorganisms operate there can be no further microbian changes, but it is probable that chemical changes not related to organisms take place. Formation of peat has been completed when microbian activity has been finally suppressed.

Humic derivatives range in color from light brown in weak concentrations to very black in high, or from the color of weak tea to strong coffee. Thiessen (1920) states that the composing materials may be differentiated into two kinds, of which one consists of somewhat heterogeneous solid particles and the other of material that appears to be homogeneous and nongranular under the ordinary microscope but granular under the ultramicroscope. The latter material is a derivative of cellulose; the former consists of macerated and comminuted plant materials, among which are spores and resinous particles. The humic derivatives contain much colloidal material, of which some have the properties of base exchange.

Such terms as ulmin, ulmic acid, humin, humic acid, ulmohumic acid, and carbohumin have been applied to the humic derivatives. These are low in hydrogen and contain about half as much oxygen as carbon. Some contain a little nitrogen. An average analysis is 63 per cent carbon, 32 per cent oxygen, and 5 per cent hydrogen.

The humins in coals have been studied by British coal chemists. The humins were extracted by alkaline solvents with or without the aid of oxidants. The extracted matter was regarded as ulmins. Those formed at the time of deposition of the organic sediments or shortly thereafter are readily soluble in alkaline solutions. These were designated *normal ulmins.* Those generated later, insoluble in primary alkaline solutions but made soluble by oxidants, were termed *regenerated ulmins.* These investigators concluded that ulmins compose the greater part of coals of high bituminous rank. A banded coal of moderate bituminous rank was found to yield 20 to 30 per cent of ulmins at 150°C. and 65 per cent at 200°C., whereas the bright jet bands, termed *vitrain,* yielded 97.5 per cent at 150°C.

The concentration of humic derivatives in organic sediments naturally varies from place to place and from level to level. The greatest concentration most likely would be beneath the surface of the deposit where leaching is least likely to be effective, where agitation of overlying waters may have little effect, where microbian activity has been brought to a close, and where the colloidal substances have been increased relatively because of dehydration. *Dopplerite* is essentially a concentrated humic derivative that perhaps was precipitated by alkaline materials in solution. This probably does not often take place. The humic derivatives permeate the preserved plant remains and ultimately serve as a binder for the organic debris.

Physiographic Conditions Controlling Deposition and Thickness of Organic Sediments

Under favorable climatic conditions, accumulation of organic sediments takes place over poorly drained areas, such as river and delta plains, glaciated areas, raised sea bottoms, subsiding areas of little relief, and lagoons and lake bottoms that have become filled and transformed into swamps. Under very favorable climatic conditions accumulation may take place on slopes of considerable inclination and even on very steep slopes.

The extensive distribution of many coal beds indicates swamps of great extent; the great thickness proves subsiding areas; the high purity of many coals shows the limited extent to which inorganic sediments could enter; and the excellent preservation of much plant structure proves that the waters had limited movement and probably were essentially stagnant.

The evidence seems to indicate that the organic sediments that made the coals were deposited on flat swampy areas of fluvial, lacustrine, or coastal plains of great extent and that many of these swamps covered areas on which soils had previously formed. The presence of stumps of trees in place shows that many swamps were more or less covered with forests. That many of the swamps were paralic and adjacent to the sea, as are the coastal swamps of the southern United States, is indicated in many by substrata of marine deposition, by the occasional intercalation of marine strata, and also by the occurrence of considerable iron sulphide in the coals of some regions, suggesting entrance of salt waters and the reduction of sulphates in solution. There was also occasional invasion of sediments on the landward margin and frequent smothering of vegetation by such. Subsidence of areas is indicated by renewal of vegetation at higher stratigraphic levels and return of accumulation of organic sediments.

Marine waters margining the places of accumulation of the organic sediments that formed some of the coals of the central United States were shallow; the seas may have been almost tideless, under which conditions a shore growth of plants could have been easily maintained and thus could have formed a screen or dam to oppose invasion of salt waters into the shoreward parts of the swamps. These shoreward parts of the shallow sea bottoms remained bathed with fresh water except during times of storms, when salt waters may have entered and remained until flushed out by flow of fresh waters from the lands.

The thickness of autochthonous organic sediments that may accumulate depends upon the extent to which water level may rise in or above the accumulating materials. The thickness of accumulation of relatively pure deposits under conditions of a generally stationary water level extends from that depth beneath water, where organic matter begins to grow abundantly, to the height to which capillary action can lift the water level. There may be limited accumulations to much greater depths, but other sediments tend to dominate. For thick accumulations an upward movement of the water level is essential. This may be brought about by clogging of drainage because of growth of plants or deposition of sediments, by damming of streams, by rise of water level in lakes because of damming of outlet, or by relative rise of sea level over a coastal swamp. The thickness of accumulation is more or less proportional to the rise of water level. To form essentially pure organic deposits of much thickness requires a nice adjustment of rise of water level to the rate of accumulation; otherwise other sediments are introduced, and the plant growth may disappear.

A third condition of the environment relates to preservation of the organic sediments after deposition. Accumulations on upland areas, as over many parts of the glaciated regions of North America, about lakes, and over floodplains are almost certain to disappear as drainage develops and lowers the water level, as outlets of lakes are deepened, and streams of floodplains deepen their channels. Accumulations in coastal and delta swamps and over subsiding areas have greater possibilities of preservation.

Climatic Conditions Controlling Deposition of Organic Sediments

Current conceptions of the layman picture the climate of the coal-forming periods as very warm and humid with great precipitation. There is no doubt that there must have been precipitation in sufficient quantity to maintain a wet substratum. Such a condition does not require great precipitation but merely precipitation well distributed. Large rainfall does not necessarily favor deposition and preservation of organic sediments but may have the opposite effect, as much water may flush the swamps of toxic products and introduce decompositional bacteria. Moderate rainfall well distributed is much more favorable for accumulation of organic sediments than much or any rainfall poorly distributed. A nice balance of precipitation is favorable. Too little precipitation leads to lowering of the water level in a swamp; too much leads to flushing and removal of toxic products.

Bearing on the quantity of precipitation required to maintain a water cover over a swamp are temperature, extent of sunshine or cloudiness, and prevalence of drying winds, as these factors influence the extent of evaporation. Less precipitation is required in regions of cold climate to develop accumulations of organic sediments than in warm. Thus, polar temperate regions with low evaporation tend to favor greater accumulation of organic matter for a given physiographic condition and a given precipitation than warmer latitudes with the same surface and precipitation, and for an equal thickness of organic matter to accumulate in warm latitudes more favorable precipitation or more favorable physiographic conditions are necessary. This amounts to stating that under average conditions more organic matter is likely to accumulate in the cooler temperate parts of the world than in the warmer parts, but that accumulation in the warmer parts is not precluded. This applies to all places except those on the margins of permanent bodies of water where conditions are essentially independent of temperature and precipitation.

Labrador with a relatively low precipitation but much cloudiness and low temperature during the growing season has great organic accumulations; Kansas with about the same precipitation but much sunshine and high temperature has essentially no organic accumulations. Organic accumulations form very rapidly in those cold regions that produce growth of plants, such as Alaska, Newfoundland, and Tierra del Fuego, and the accumulations may even form on steep slopes, tops of crags, etc., as the coolness retards and may even suppress microbian activity. But organic accumulations also form in tropical regions of low relief in spite of the fact that microbian activity is favored and ordinarily is high. This is due to a smothering plant growth under conditions of a well-distributed precipitation that need not be particularly high but is adequate to maintain a relatively permanent plant cover.

The plants of the coals suggest that climatic conditions were generally equable, with moisture adequate to maintain a wet or water-covered substratum. It may be assumed that the climatic conditions were rarely severe, but the presence of tree rings in some of the plants of the coals deposited in temperate and subpolar latitudes shows seasonal cold. Equability of climate is also shown by the extensive distribution of identical or similar plant associations. Any generalization respecting the climate existing at the place and time a particular coal bed was formed should be made on the basis of the plants in the coal and the sedimentational features in the coal and associated sediments that reflect climate.

Effects of Changes of Water Level and Drainage

There can be little doubt that water levels in the coal swamps changed from time to time; otherwise the organic accumulations would have been very much the same from bottom to top. This is not the case. Change of water level may have been done by (*a*) erosional deepening of outlets of the swamps; (*b*) obstruction to escape of waters either by growth of plants in the outlets, enlargement or decrease of the barriers to which the swamps may have been due, deformation raising or lowering the barriers, or to stream or current waters damming back the swamps' runoff; (*c*) capillary rise or fall of water in the accumulating sediments; or (*d*) climatic or weather changes affecting precipitation, temperature, evaporation, or plant growth. Some of the factors were of seasonal application; others applied over terms of years; and others were operative over long time periods. But each probably produced changes in the rates of growth and deposition of organic sediments and microbian decomposition and thus differences in the character of the deposits entombed. Some of the changes of level led to introduction of inorganic sediments and the formation of "breaks" or "bone" beds. Great concentration of toxic products was made at times, and there was correspondingly great preservation of cellulose, and the coals are humic. Or there was thinning of the toxic products, more microbian decomposition, more elimination of cellulosic materials, and relative increase in bituminous constituents. The water level may have fallen at times to expose the surface of the organic sediments with the formation of thin layers of cracked organic material, and there may have been some leaching.

Environments Indicated by Plants in Coals

The plants of autochthonous coals had adaptive structures like those of living plants that dwell in the swamp environment. Among these are the swollen and dilated basal parts of the trunks of *Sigillaria*, *Lepidodendron*, and some of the tree ferns; wide-spreading and shallowly distributed roots, as grow on living swamp trees; bark with air pores, root stems, and rhizomes at successive levels as in *Calamodendron*, *Annularia*, and others;

thin-walled laciniose tissue in some of the cortical cylinders of the giant lycopods and tree ferns; provision for the storage of water as in *Sigillaria*, *Neuropteris*, and others; various pseudo-xerophytic structures; flotation chambers in some of the Paleozoic fernlike higher plants; a heterosporous type of reproduction in many trees; water-resistant coverings of seeds, spores, and pollen of many plants; air chambers in calamarian roots and Micorrhiza on the roots of *Calamites* for growth in peat.

Substrata of Coal Beds

The substrata of coal beds, if composed of clays or mixtures of clay and coarser constituents, are known as underclays. The top parts of an underclay are generally without stratification, often appear as if leached, very commonly contain roots, and in many instances have stumps of trees and smaller plants in positions of growth. It is obvious that in many instances the underclays of coals represent old soils. But not all substrata are old soils. Some are beds of sandstones, shales, and even calcareous rocks.

In some cases the surfaces upon which coals rest bear evidence of erosion, after which plants grew and organic sediments accumulated. In other cases the surfaces are those of deposition, upon which the organic matter grew and accumulated or was deposited after transportation; in the latter case the organic sediments are allochthonous. Allochthonous sediments tend to have limited and variable thickness and small horizontal extension, are rarely composed of very pure organic matter, and may rest upon any kind of inorganic sediment to which they are sequential. (The cannels are not included in this statement.)

CLASSIFICATION OF CARBONACEOUS SEDIMENTS

Introduction

The carbonaceous sediments may be placed in the two classes of *humic* (*coaly*) and *sapropelic* (*bituminous*). There are all gradations from one to the other. The humic (coaly) sediments are low in hydrogen and high in carbon and oxygen, whereas the sapropelic are high in hydrogen. The former is exemplified by the ordinary coals and the latter by the true cannels. Further division is possible in each variety, but places of division are not easy to select. Another factor entering into the matter of classification arises from the mingling of organic and inorganic sediments in all possible combinations.

Basis for the differentiation of the different kinds of organic sediments rests to some degree on differences in the kinds of plants but to a greater degree on the extent of microbian action before suppression. In some cases the organic sediments are from land plants, in others from aquatic plants. Accumulation in some cases takes place in very stagnant and very toxic waters, with little decomposition because of early suppression of microbian action. In other cases microbian action accomplishes great elimination of carbohydrates.

Although plant materials seem to have been composed of protein, cellulose, carbohydrate, lignin, fat, oil, wax, gum, and resin from the beginning of known geologic history, the various substances are not now in the same proportions in different plants, and the proportions probably have varied from time to time. Certain parts of plants, as pollen, spores, some bark, etc., are high in fatty, waxy, and resinous substances, whereas other parts contain little of these materials. Selective microbian action tends to bring about relative increase of bituminous constituents. With several variables making possible many gradations between the extremes of organic sediments that are finally entombed, there is naturally difficulty of classification of the original organic sediments and equal difficulty of classification of the lithified equivalents.

Stopes (1919) classified coal into four rock types: *vitrain*, *clarain*, *durain*, and *fusain*. Each was stated to be composed of substances termed *macerals* such as *vitrainite*, *fusinite*, *exinite* (spore exines, cuticles, etc.), and *micronite* (a granular, colloidal coaly substance, opaque in transmitted light). These have been discussed by Dapples. Cady states that coals are composed of *phyterals* (particles based on recognition of plant fossil parts and pieces) and *macerals* (Stopes, 1935) (defined as entities recognized by virtue of chemical and physical similarity). The two entities are not mutually exclusive in that phyterals may be represented by several types of macerals, and a maceral may be represented by several phyterals. Any variety of coal is probably a mixture of several phyterals and macerals.

Thiessen (1931) states that coals are composed of two different classes of constituents which he terms *anthraxylon* and *attritus*. The unlithified equivalents of these are in the original organic sediments.

Anthraxylon is derived largely from the visible parts of plants, as leaves, stems, roots, and bark, all more or less broken prior to coalification. These were impregnated to some degree with humic derivatives and may be seen more or less flattened in coal as bright, glossy, or jetlike bands or lenses. These are known as *vitrain*. It is thought by Thiessen that anthraxylon of all coals is largely derived from lignin.

Attritus consists of very finely divided organic matter that is composed of the more resistant parts of plants and secondary products resulting from microbian action. It may originally have been high in finely divided carbohydrate matter at one extreme and in fatty, waxy, and resinous materials at the other. Thiessen differentiates two kinds: *lucid attrite* and *opaque attrite*. Lucid attrite is distinctly transparent in thin section; opaque attrite is opaque in ordinary thin section, but becomes translucent in very thin section. Concentrated in lenses and bands, attritus is known as *durain*.

Fusain, the so-called *mineral charcoal*, or *mother-of-coal*, is present in many coals (Fig. 50). This has been interpreted as charcoal due to fires over the peat swamps or elsewhere, but the evidence is not in accord with this interpretation. David White (1932) states that the evidence indicates that fusain represents fragmental organic matter impregnated and coated with humic derivative. Evaporation of the water cover of a swamp rendered the humic concentrate insoluble and thus resistant to collapse on being wetted. Fusain is higher in carbon and lower in volatile matter than most components of coal, and it has few coking qualities. It resembles charcoal and may be seen over bedding planes.

Coals composed largely of anthraxylous or woody matter are very common, very fragile, and have better coking qualities than attrital coals of the same rank. They are the bright or glance coals. Anthraxylous coals have high humic and low ash content, general rarity of fusain or the so-called mineral charcoal, and more or less excellent preservation of organic structures. Coals very largely composed of attritus have a more dull appearance than anthraxylous coals. The two constituents of coals may be more or less intermingled with each other or be arranged in thick or thin bands, and the bands in association may be of varied thickness. These are the banded or striped coals, termed *clarain* by British students. Coals composed very largely of attritus with anthraxylon largely absent may be differentiated into *cannel coals* composed of spores and similar organic materials, *humic* or *pseudocannels* largely composed of humic matter, and *boghead cannels*, of which fatty algae are the chief ingredients (Fig. 51). The true cannels are considered under the head of bituminous sediments.

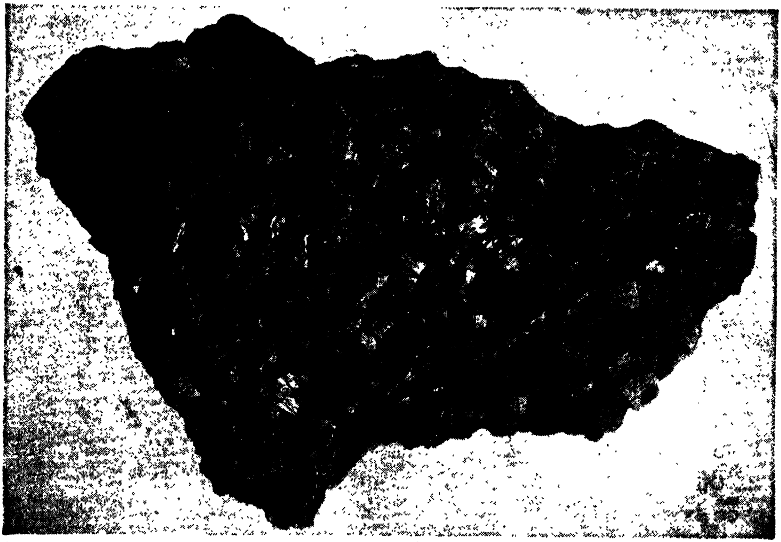

FIG. 50. Slab of coal showing fusain (× ⅔). Fragments moved from original positions and then immersed in all positions in a carbonaceous matrix. Shows evidence of some disturbance and redeposition of carbonized organic matter.

The original organic sediments from which the carbonaceous sediments formed may be placed in the three groups of peat (either anthraxylous or attrital, or both), *muck*, or *sludge*, and *sapropel*. There are all gradations.

Peats

In conformity with the terms used by Thiessen, peats may be placed in the two classes of anthraxylous and attrital. Anthraxylous peat is dominantly composed of little-modified original plant materials that are more or less permeated with humic derivatives.

As the extent of microbian action varies from year to year, it results that some levels are high in anthraxylous matter and others have a large content of attritus. This results in some degree of stratification. This is the most common type of peat. Coalification produces a banded or striped coal of which the bright jetlike layers or laminations, originally anthraxylous matter, are composed of anthraxylon (vitrain) and the dull mat layers of attritus (durain). If decomposition is limited and only the little resistant materials are eliminated and conditions over the swamp are essentially uniform from year to year, the peat is largely composed of lignocelluloses and the more resistant plant materials and is without stratification. This forms anthraxylous or vitrainous coal. If decomposition proceeds to extent of elimination of all, or nearly all, carbohydrate materials, the sediments become bituminous and approach the sapropelic deposits and result in cannel coals. There are gradations between the two types of peat, and each may pass laterally into the other. Lenses and layers of cannel are common in anthraxylous coals,

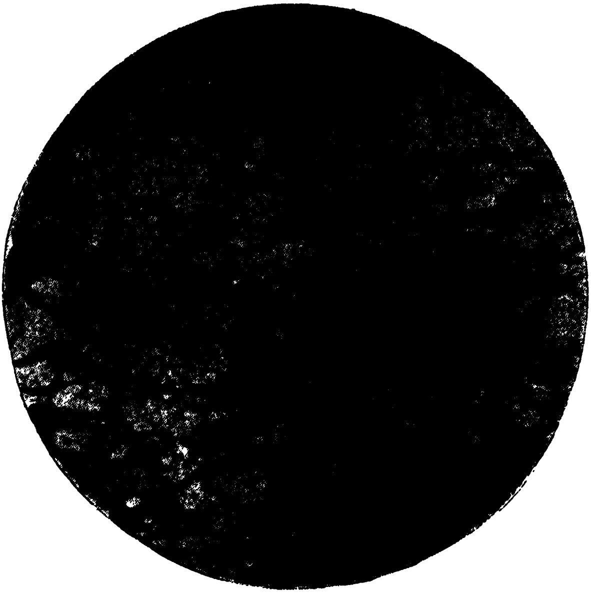

FIG. 51. Section of torbanite (algal boghead) parallel to bedding (× 200). Shows the algal colonies of *Pila scotica*. From Linlithgoshire, Scotland. (*Photograph by R. Thiessen.*)

and, similarly, lenses and layers of anthraxylous coals are present in cannel coals.

Submergence or Emergence of Peat Deposits. Falling of the water level over a peat swamp exposes the organic deposits to destructive agencies, and if left unprotected sufficiently long, the deposits totally disappear. Rising of water level may permit more agitation, greater aeration, additional decomposition, a probably greater introduction of inorganic sediments, and possible erosion of some of the organic sediments already deposited. Some loss is probable, and, if deposition of inorganic sediments does not immediately replace organic deposition, or if organic deposition does not continue, the loss may be large and the whole of an organic deposit may be destroyed. If inorganic sediments are deposited, there may be a transitional thickness of mixed organic and inorganic sediments, or the inorganic sediments may directly follow the organic. If entering waters are salt or brackish or contain sulphates in solution, there is likely to be formation of iron sulphides in the upper part of the organic sediments, in the transitional strata, and in the lower part of the inorganic sediments, particularly if the conditions were such as not to permit aeration of the waters. Deposition of inorganic sediments ultimately protects the organic sediments against attack but also builds up the bottom so that ultimately a depth is attained that may permit renewal of swamp conditions.

Rate of Accumulation of Peat. Owing to the various factors influencing accumulation of peat, among which are the rate of production of organic sediments and microbian action, it is probable that there are as many rates of accumulation as there are combinations of conditions. Data derived from the glaciated regions of the world suggest that the maximum rate of accumulation is about 0.3 meter in 10 years, which, under favorable conditions of preservation at the present time is thought to produce 10 to 12.5 cm. of peat by the time it has been buried to the depth of 6 to 7.5 meters. David White (1932) suggests that 1 ft. (0.3 meter) in 75 years at the depth of 25 ft. (7.5 meters) is a conservative average, and perhaps 1 ft. of dense hard peat may represent 100 years. Ashley suggests that 1 ft. at the surface will shrink to 3 in. (7.5 cm.) owing to loss of moisture and that in a deep bog this will decrease to 1⅛ in. (2.8 cm.). It is possible that under other conditions of growth and with other vegetation the rate of accumulation may have been greater. Consideration of the thickness of peat necessary to form a foot of coal is fraught with great possibility of error, as the losses of water, hydrogen, carbon, and nitrogen are subject to such great variations. Renault concluded from study of the coalified stems of plants in coals of bituminous rank that the volume had been reduced to $\frac{1}{17}$ to $\frac{1}{20}$ of the original, which, if applied to raw peat, would require about 6 meters of raw peat to produce 0.3 meter of coal of bituminous rank. This gives no consideration to losses due to microbian action and removal of humic derivatives from swamps. Ashley estimated 2,100 years as the time required to deposit the organic matter of the Pittsburgh coal bed, where it is 7 ft. (2.1 meters) thick, or 300 years as the time required to deposit the vegetable matter for 1 ft. of coal. This figure probably errs in allowing too short time.

The Ash of Coals

This topic applies to ordinary or humic coals and does not include the true cannels. The inorganic or ashy constituents of the carbonaceous sediments were in part derived from the organic materials themselves and in part from inorganic substances in solution and suspension in water flowing into a swamp on in suspension in air passing over a swamp. The quantity of materials not derived from the plants themselves is subject to wide latitude. It is thought that 2 to 6 per cent may be considered to have had origin in the inorganic constituents of the original plant matter, although that percentage of ash is high for some plants, and that ash in excess of these figures was brought to the swamps by wind or water or both. Lessing found that average percentages of ash in the various constituents of coal are as follows: fusain, 15.59; durain (opaque attrite), 6.26; clarain (partly vitrain, partly durain, and some fusain), 1.22; vitrain, 1.11. It is obvious that components least modified by microbian action, for example, vitrain, should have the least ash.

Bone Coal

All forms of humic sediments contain some inorganic material. If the quantity is large, the aggregate is known as muck, or sludge, which is defined as a deposit of inorganic sediments with a more or less high content of carbohydrate matter in distinction to sapropelic sediments in which the organic matter is high in bituminous ingredients. Colors tend to be black, but they may be brown if exposure has permitted oxidation of iron. "Bone" or "bone coal" is lithified muck. Layers of "bone" in coal beds represent times when considerable quantities of inorganic sediments were brought to the swamps and mingled to a greater or less extent with the organic sediments. In some cases there are layers of essentially inorganic sediments with only small quantities of organic materials, and in others the inorganic sediments carry enough organic matter to form a low-grade fuel. "Bone" frequently contains excellently preserved plant remains. The waters that brought the inorganic sediments may have come from the sea or from the land and may have been salty, fresh, or of some intermediate concentration.

Sapropelic Deposits

Sapropelic deposits consist of muds, slimes, and oozes that are high in bituminous constituents and are known as *sapropel*, *gyttja*, or *dy*. They were deposited in more or less open waters. There is great variation in composition. The organic materials may be largely composed of the more resistant parts of plant structures, but if the waters of deposition were sufficiently toxic, some of the original plant structure may have been preserved. Woodland lakes of northern Wisconsin contain extensive deposits of gyttja of which the chief organic constituent is a pale greenish-yellow gel (Twenhofel, McKelvey, Carter, and Nelson). Sapropelic deposits tend to be allochthonous.

Deposition takes place in open waters relatively free from inorganic sediments, as ponds and sloughs in swamps, shallow lakes, inland seas, and bays, gulfs, and sounds of bodies of water with weak tides, waves, and currents. The waters should be protected against entrance of much woody matter and should be too deep or too toxic to permit much growth of peat-forming plants. Conditions, however, should be such as to permit microbian action to proceed to a considerable degree before suppression, so that anthraxylous matter largely disappears. The organic sediments deposited in such waters either grow there, or the plant matter is of terrestrial growth and is brought in by water and wind. Adequate aeration insures microbian action being carried nearly to its subaqueous limit and disappearance of unprotected cellulosic and ligneous materials, thus leaving only such substances as resistant resins, waxes, fat, and gums and cellulosic and ligneous materials impregnated with toxic derivatives. If the open waters are suitably situated with respect to terrestrial plants, they may receive more or less large quantities of pollen grains and spores, and thus the deposits become high in these substances. Other parts of terrestrial plants may enter in quantity and variety dependent upon the climatic and physiographic characters of the regions and the relations of the dimensions, shapes, and positions of the water bodies to the regions in which they are. Remains of fatty and waxy algae may be rare to abundant, and in some cases they compose essentially the whole of a deposit. Species of *Pila* and *Reinchia* seem to be abundantly represented, as shown by Thiessen in his studies of the algae in the recent coorongite from the Coorong District of Australia.

Hard parts of animals may be present in sapropelic deposits, but remains of benthonic organisms should not be common, as bottoms of sapropelic accumulation are not favorable dwelling places for benthos. The toxic waters over bottoms of sapropelic accumulation may entirely suppress benthonic organisms, and, should conditions permit their presence, the shells would likely be somewhat smaller than those of animals living under more congenial conditions. Burrowers and organisms eating the bottom for the contained organic contents may also be present, and the sediments thus may make one or more passages through intestinal tracts and be in some degree of excremental origin. Preservable parts of planktonic and nektonic animals of overlying waters may be common to abundant in sapropelic deposits in the forms of bones, shells, fish scales, etc. There is no proved record of soft animal tissues, but it can hardly be doubted that derivative products of such tissues are present. Renault has stated the probable survival of insect eggs in a peat bog and their silification in advance of compression, and what seems to have been fat was obtained in a single dredging from a position in the Atlantic Ocean about 5 deg. west of the Spanish Coast (Gümbel). E. T. Erickson has stated his belief that some degree of preservation of animal tissue may take place in sufficiently toxic solutions containing humic acid and thus be preserved beyond the biochemical stage and become available for later generation of hydrocarbons.

Waters that contain sulphates in solution may become charged with hydrogen sulphide and ultimately sulphuric acid because of activity of sulphate-reducing bacteria. The sulphuric acid would lead to disappearance of carbonate shells and cause deposition of iron sulphides in sapropelic deposits. Sulphides would not be present in waters free from sulphates, unless some of the organic matter contained sulphur.

If sapropelic sediments accumulate under conditions excluding inorganic sediments, the product after lithification is a high-grade cannel coal (Fig. 52). If considerable quantities of inorganic sediments are introduced, the deposits form a black or oil ("cannel," "bastard") shale. If the waters are such as to favor growth of fatty or waxy algae, the deposits are distinctly aquatic and bituminous from the beginning and form boghead cannels or boghead cannel shales. As sapropelic sediments are deposited in open waters, they are likely to be stratified to a greater or less degree. Cannel coals and shales that seem massive on fresh surfaces are very apt to separate into thin laminae after some exposure.

There are all gradations between sapropelic deposits composed dominantly of spores and kindred substances that are high in fatty, waxy, and resinous materials and the boghead algal sapropels largely composed of the remains of fatty algae. Likewise there are all gradations between sapropelic and humic deposits.

Lake waters of the temperate zones do not offer the most favorable conditions for preservation of sapropelic sediments, because of their semiannual overturns, which bring oxygen to the bottoms twice each year and at the same time remove toxic products from the bottom. This permits aerobic bacterial activity that may lead to complete elimination of organic matter. Waters of tropical regions have limited overturn, and accordingly bottom waters retain toxicity, and destruction of organic matter is not so likely. Organic matter may be very excellently preserved in the lake waters of very cold regions, the cold bottom waters retarding and perhaps ultimately suppressing microbian decomposition, thus leading to preservation of much organic matter that in more temperate regions would be eliminated.

Sapropelic deposits have been studied by Potonie, Stremme, Späte, and others. Trask has examined sapropelic sediments from many parts of the world. On a small scale they may be seen on bottoms in the open waters over most swamps and muskegs of all parts of the world. They have extensive occurrence in the sounds and bays (limans) on the east coast of the almost tideless Baltic Sea, the landlocked fiords and bays of Norway (Strøm), the barred basins of Woolnough, the more or less stagnant depths of the Black Sea, many fresh-water lakes, and many saline lakes of desert areas of the world, and they may be expected to have had extensive occurrence in some of the more or less tideless landlocked epeiric seas of the past.

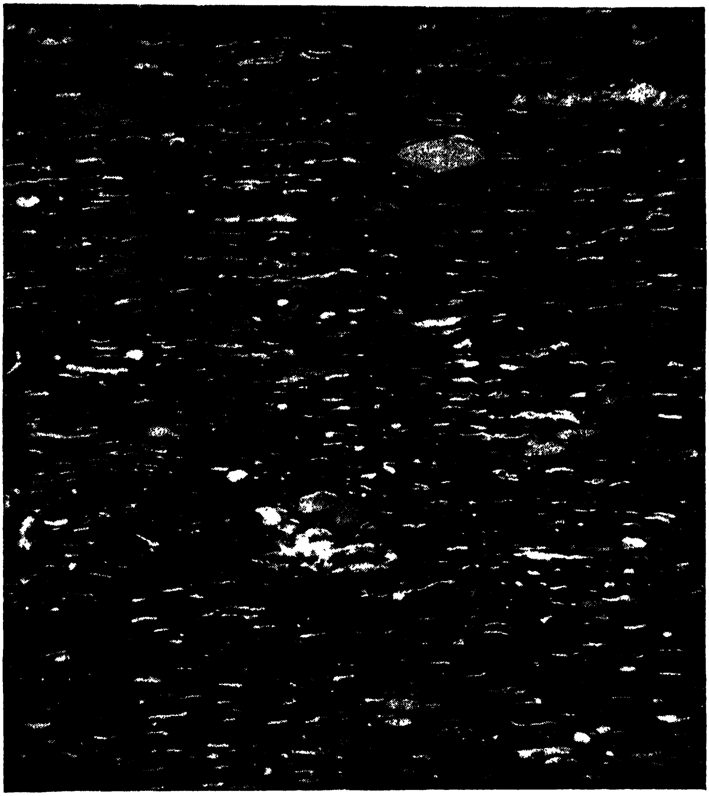

FIG. 52. Vertical section of cannel coal from the Upper Kittaning Bed at North Washington, Penn. (× 160). Shows a fatty stratum composed mainly of concentrated spore exines, showing as irregularly lenticular bodies; resin lumps; and angular fragments of cuticle in the form of thin, irregular horizontal elements of varying lengths. All of these substances are immersed in the dark humic derivative ground mass. (*Photograph by R. Thiessen.*)

Varieties of Sapropelic Deposits. Lithified sapropelic deposits range from those containing only small quantities of organic matter to others containing only small quantities of inorganic matter. They may be placed in two classes on this account, one dominantly of organic ingredients and the other with a limited content of organic ingredients. The former is represented by cannel coals, the latter by the black or oil shales. Some of the latter may be highly calcareous and approach limestones.

As previously noted, black shales are of two kinds, of which one has organic matter largely of terrestrial origin and composed of carbohydrate, cellulosic, and ligneous materials, and the other has organic matter largely of aquatic origin and composed of spores, algae, and other fatty, waxy, and resinous materials, deposited in open waters. The former is associated with swamp conditions and coal beds; the latter is to a considerable extent marine and at any rate is distinctly aquatic. The former has been considered in other connections; the latter are the oil shales. The extent of blackness of oil shales is some measure of the organic content but not completely so. The plant matter may be excellently preserved with little maceration, or it may have been so mutilated as to be unrecognizable. Microbian action may have been of limited application, or it may have been carried well toward completion. There are all gradations.

Many sapropelic shales contain marine fossils (see topic Black Shales). These are not commonly benthos, but mainly plankton and nekton, giving evidence that these sediments were deposited in connection with marine waters.

Between oil shales and bituminous black shales there are no differences, as oil may be obtained from both. Shales are termed oil shales if they yield oil in considerable quantity on distillation. The organic matter is distinctly bituminous and consists of spores, spore cases, pollen grains, or fatty and waxy algae. These are immersed in the products of animal and plant microbian decomposition. Some oil shales are filled with colonies of algae of which the structures are often excellently preserved. Colors may be black, as in the Chattanooga shale of the Ohio Valley and the kerosene shale of New South Wales; brownish black, as in the Scottish torbanite; brown, as in shales from Brazil; buff, as in the "Oil Rock" in the Platteville of Wisconsin; and nearly red, as in the Braunschiefer of Estonia.

Some seem massive, but split into laminae on weathering. In many the laminae are paper-thin and frequently somewhat leathery and elastic.

The organic matter of oil shales seems to be mainly noncrystalline, but E. S. Larsen has found minute birefringent crystals in the organic matter shown in thin section of oil shales, and Bradley observed several varieties of crystals in organic matter of the Green River shales of the northern Rocky Mountain Region of the United States.

The richest oil shales are the bogheads. These seem usually to be composed largely of a single kind of plant, and this may be present in excellent preservation. The *kerosene oil shale* of New South Wales is one of the richest in oil of the bogheads. The organic matter of this shale is largely composed of colonies of the colonial alga *Reinchia australis*, of which the cell walls are lined with a waxy deposit. Boghead shales in the Permian of Autun, France, have a somewhat similar composition, but the plant matter consists largely of colonies of *Pila bibractensis*. In the Scottish oil shales the plant matter seems dominantly composed of *Pila scotica*. The organic matter of the Esthonian boghead is composed of a unicellular alga closely resembling the living *Glaeocapsa*. *Closterite* is a dense, laminated, brownish-red canneloid material from the Irkutsk River Basin of Siberia, of which the organic matter is composed of spheroid algal colonies of different sizes, which in some respects are similar to those of *Protococcus botryoides*. There are also remains of a desmid alga belonging to *Closterium*, a living genus. Many of the remains are excellently preserved. *Tcheremkhite* is an algal sapropelic deposit found in the vicinity of Cheremkhovo, Russia, which has been interpreted as an aggregation of peaty matter washed from other deposits. It has yellowish to brownish color and contains small reddish-brown pellets of various forms and yellow cellular structures closely resembling those of *Pila*.

Lamination of oil shales may be seasonal as postulated by Bradley for the Green River shales (Fig. 53), or due in some cases to generative periods of the organisms responsible for the organic matter. Features resembling desiccation breccia are present in the Green River shales and are referred to exposure and mud cracking. This indicates shallow waters and their occasional evaporation, leading to concentration of dissolved salts and development of high salinity. Some beds in the shales suggest that saline efflorescence has taken place. Shallowness is also indicated in the Green River shales by lenticularity of bedding, but the waters seem generally to have been too shallow or the sediments too fine-grained to develop cross lamination, since it is rare. Fossil remains, particularly benthonic, are rare, but remains of fish are not uncommon. Skins of larvae of different types are rather widespread, and some layers in northwestern Colorado and northeastern Utah are more or less covered with them. The small dimensions of the leaves contained in the shales suggest aridity of climate.

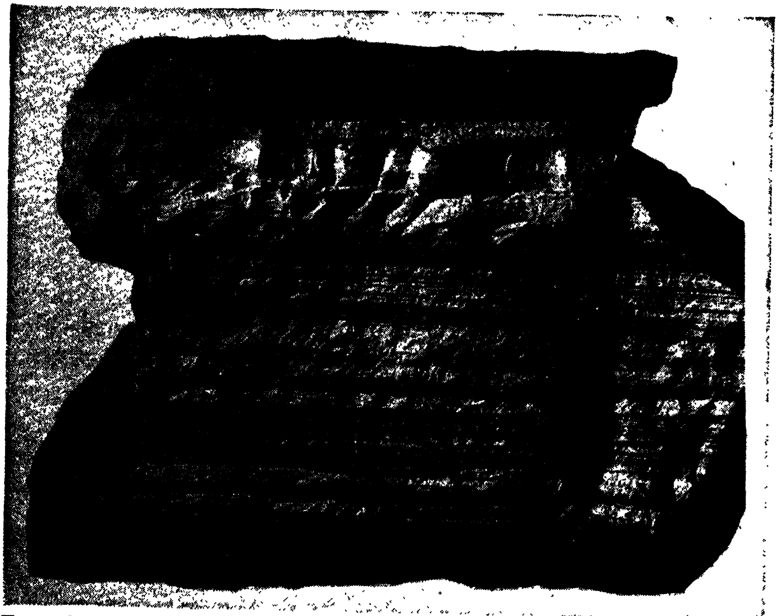

FIG. 53. Vertical section of Green River shale (× ¾). (*U.S. Geol. Surv.*)

Several recent algal sapropels have been subjected to considerable study, and these studies aid in interpretation of fossil occurrences. Among recent sapropels is *coorongite.* This originates as a greenish scum on lakes in the Coorong District of South Australia. Evaporation of the waters spreads the scum over exposed bottoms and shores in the form of a thick, brownish, more or less elastic blanket. It has been found to consist mainly of unicellular algae that are much like *Reinchia* of the kerosene oil shale of Australia and *Pila* of the torbanite of Scotland. Coorongite is very rich in bituminous matter. Associated inorganic sediments are mainly sands. Analysis of coorongite in the laboratory of the United States Bureau of Mines gave results as follows:

TABLE 25

Constituent	Per Cent
Moisture	1.6
Volatile matter	90.1
Fixed carbon	2.6
Ash	5.7
Total	100.0
Hydrogen	11.3
Carbon	73.8
Nitrogen	0.7
Oxygen	8.4

N'hangellite is a somewhat similar substance from Portuguese East Africa, where it forms crusts several inches thick and is used by the natives for fuel. Similar algal accumulations composed of *Botryococcus brauni* are forming in the region of Balkhash Lake in Turkestan, and the deposits are stated to be as much as 12 meters in thickness. The algal materials are blown or washed to the shores or become stranded on the bottom to form elastic yellow mantles on evaporation of the water.

It is likely that organic deposits similar to coorongite are common over many parts of the world and that they represent oil shale like torbanite and the kerosene oil shale in the making. Those described are from semiarid to arid regions, and it is possible that aridity may be a favorable but not a necessary factor, as deposits rich in bituminous constituents are known to form in humid environments.

GEOCHEMICAL CHANGES IN ORGANIC SEDIMENTS SUBSEQUENT TO DEPOSITION

After burial of organic sediments, they are subjected to gradually increasing pressures, compacting, and a rise in temperature. These are physical processes, and various physical and chemical changes have been attributed to them. The processes lead to dehydration, increase in density, and formation and expulsion of volatile substances. These changes have been designated *dynamochemical* or *geochemical,* and through them advance in the rank of an organic deposit is made from peat to lignite (Fig. 54), thence to bituminous coal (Fig. 55), thence to anthracite, and finally to graphite. Not a great deal is known respecting details of the changes that take place in the range from peaty and sapropelic deposits to coal and hydrocarbons. At some stage in the metamorphism wood ceases to be cellulose and lignin, and there occurs a change from waxes, fats, gums, and resins to hydrocarbons or substances from which hydrocarbons may be distilled.

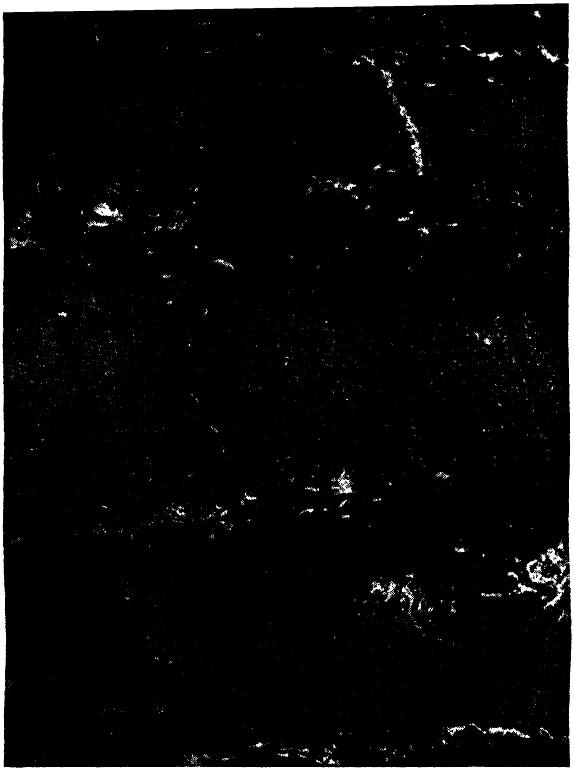

FIG. 54. Vertical section of lignite of Fort Union age from Wilton, North Dakota (× 200). Three bands of wood cells are shown, and wreckage of cells fill the separating zones. (*Photograph by R. Thiessen.*)

Many attempts have been made in the laboratory to solve some of the problems connected with the changes noted above. These have followed two general lines, of which one has attempted to form coal and the other to generate hydrocarbons that could be removed by solvents.

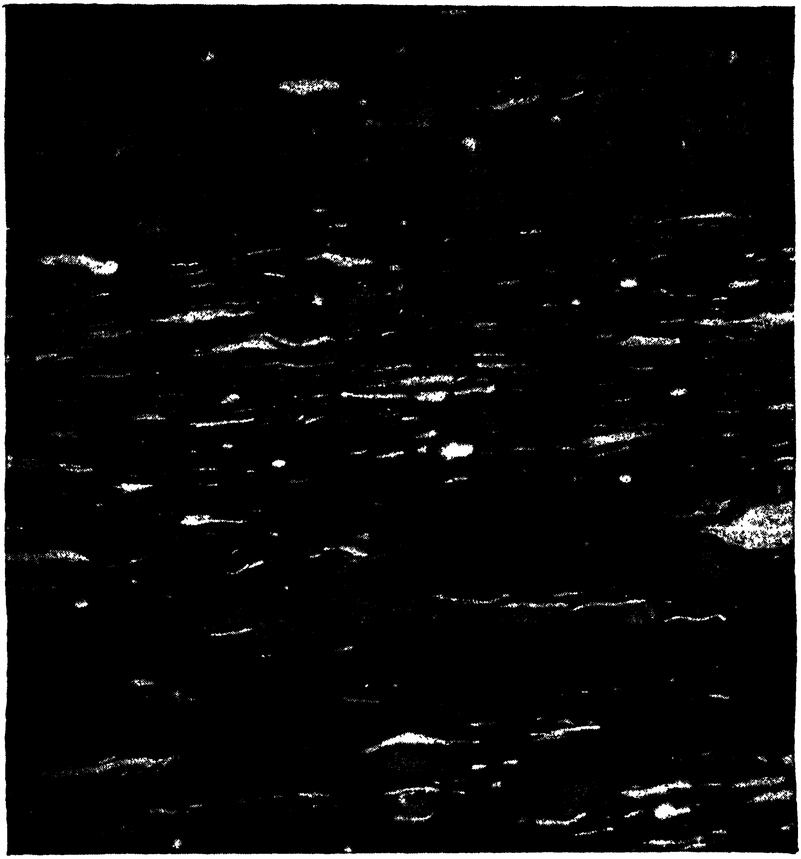

FIG. 55. Vertical section of coal from Brookville Bed, Grove City, Penn. (× 200.) More decomposed material is shown at the bottom. Above this and again near the upper part are zones of variable decomposition. Lumps of resin are shown in the central part. (*Photograph by R. Thiessen.*)

Formation of coal has been attempted many times. Spring submitted peat to a pressure of 6,000 atmospheres and reported formation of a brilliant black solid resembling coal. Attempts by Zieller to confirm his results did not succeed. Zieller found that peat and ulmic acid after submission to pressures ranging from 2,000 to 6,000 kg. per square centimeter remained unchanged, and Gümbel found that lignite after being subjected to pressures of 20,000 atmospheres was still lignite. Coal is said to have been made by heating wood in a closed cylinder or tube, and Fremy secured products resembling low-rank coals by heating carbohydrates and ulmic acid extracted from peat to 300°C. in sealed tubes. Stein after heating wood and water in sealed tubes to temperatures of 245 to 290°C. obtained coaly materials which, depending on the temperatures used, ranged from materials resembling peat to those resembling ordinary coal.

Results of experiments by Taylor led to the view that the base-exchange properties of the sediments overlying coals (roof of coal) have much to do with the rank of an organic deposit, and he formulated the following generalizations:

1. Acid conditions without a roof result in peat.
2. Acid conditions under a roof containing a calcium aluminosilicic complex result in lignite.
3. Anaerobic conditions under a roof containing a sodium aluminosilicic complex result in bituminous or anthracite coal.

Additional data are necessary before these generalizations can be considered more than hypotheses.

The various experiments and observations that have been made permit the generalization that pressure, temperature, and time are important factors in bringing about the transformation of peats to coals. Temperatures applied were not great, and, given longer time intervals than those used, it seems likely that much lower temperatures might be equally effective.

It is well known that most bituminous sediments yield greater quantities of hydrocarbons after heating than can be extracted by solvents. It has been suggested that the bituminous materials are in the form of a substance termed *kerogen* which may be transformed into soluble hydrocarbons by heating. The composition of the postulated kerogen is not known (Rae).

Attempts have been made to determine if shearing pressures have anything to do with the transformation of organic matter in the cannels and oil shales into petroleum. Results have not been wholly in accord, even when the work has been done by the same investigators (Hawley, Rand).

The original organic sediments are composed of carbon, hydro-

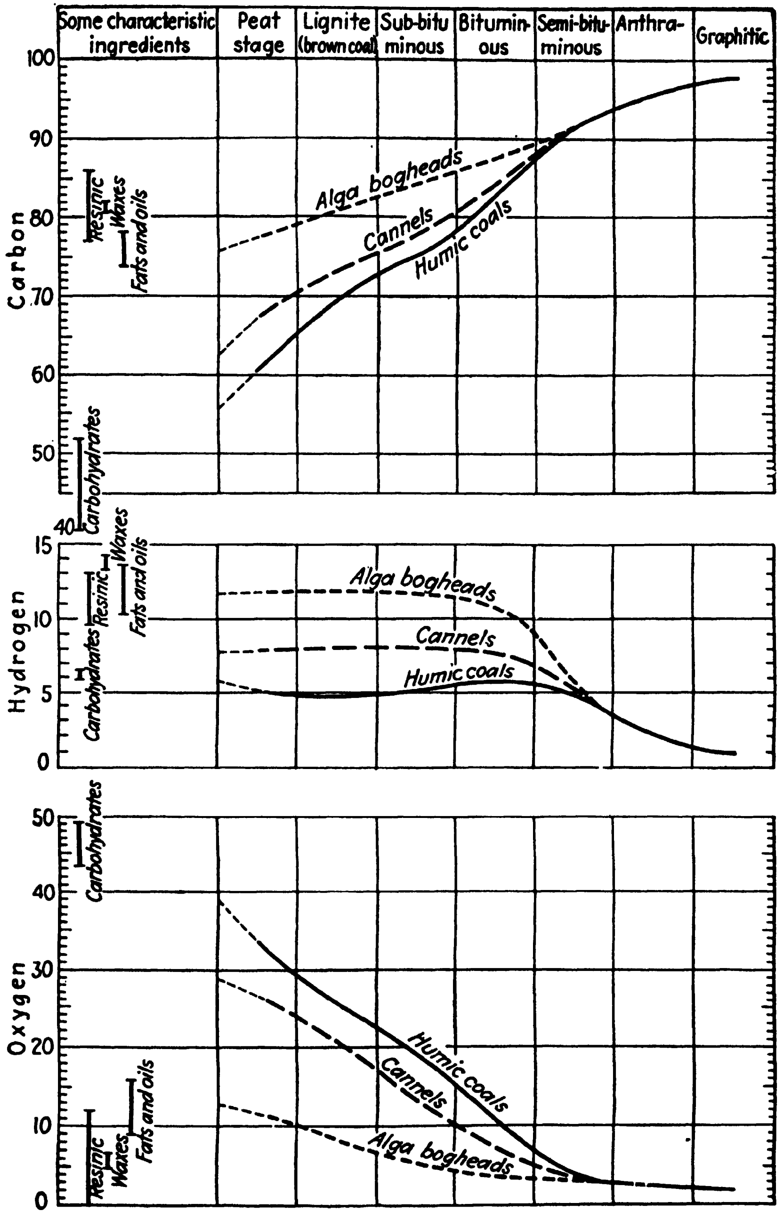

FIG. 56. Diagram showing relations of carbon, hydrogen, and oxygen in humic, cannel, and alga boghead coals in the advance from low to high rank. All approach 100 per cent carbon and zero oxygen as a limit when the ash is disregarded. [*After David White, Progressive regional carbonization of coals, Am.*

gen, and oxygen. Alga bogheads have the highest carbon and hydrogen content and the lowest oxygen content; the sediments ultimately resulting in the humic coals have the lowest carbon and hydrogen content and highest oxygen content; and sapropelic sediments ultimately resulting in ordinary cannels and ordinary oil shales have contents of these three elements intermediate between the humic sediments and alga bogheads. As these respective sediments advance in rank by reason of dynamochemical metamorphism, they all produce products with a high but similar carbon content and a very low but also similar oxygen and hydrogen content. This is shown in the diagram of Fig. 56.

SECONDARY CARBONACEOUS MATTER

The transformation of coaly and bituminous substances through dynamochemical processes is attended by the production of various secondary substances, which are removed in various ways from the sediments and may accumulate elsewhere in the rocks. These are various compounds of carbon, oxygen, hydrogen, nitrogen, and sulphur. The gaseous compounds form natural gas, which is commonly CH_4 and CO_2. Liquid by-products are the natural oils, and there are also some solid compounds of carbon and hydrogen.

Coaly substances yield volatile substances, which are water, carbon dioxide, gaseous hydrocarbons, apparently tarry hydrocarbons, and perhaps liquid hydrocarbons in small quantity. Lignite averages more than one-half volatile materials. Humic coals of bituminous rank have a volatile content of 47 to 25 per cent; those of semibituminous rank, 25 to 17 per cent; and anthracite from 7 per cent to less.

The volatile matter in bituminous sediments is a much higher percentage of the combustible matter than in the coaly peats, and in the purest alga bogheads it may exceed 85 per cent. Oxygen falls under 15 per cent in the peat stage of the bogheads and under 4 per cent after they have advanced to bituminous rank. The diagram of Fig. 56 shows the losses of volatile matter from a quantitative point of view and the evolved liquids and gases attendant upon the dynamochemical processes.

As elsewhere, the time factor is important in the transformation of organic matter into coal and oil. Studies by Maier and Zimmerley of Green River shale enclosed in sealed pyrex tubes showed that a much lower temperature is required for the transformation of organic matter into petroleum if the heat is applied for relatively long intervals than if applied for short intervals. They erected the formula of $T = 8.4 \times 10^5$ years for the time required for 1 per cent transformation into hydrocarbons of the bituminous matter of oil shales. Trask corrected the formula to $T = 8.4 \times 10^4$ years. It has been found that in most high-rank bituminous coals no extensive decomposition and no yield of hydrocarbons take place below 300°C., and the distillation of considerable hydrocarbons takes place in most woody bituminous coals between 310 and 350°C. Decomposition of "ulmin compounds" begins above 300°C., and production of oils from the ulmin compounds is confined largely to a temperature range of 25 to 30°C. above the "active decomposition point."

The coaly sediments are generally considered to yield limited quantities of hydrocarbons which are mainly gases with some tarry substances and a limited quantity of liquid hydrocarbons which are said not to be particularly typical of petroleum. The sapropelic sediments with their large content of wax, fat, gum, and resin are those that may be looked upon as the chief source rocks from which the liquid hydrocarbons were derived, and they are thought to yield these in quantities somewhat proportional to the original quantity of bituminous constituents present, the alga bogheads being considered the most prolific in yield.

It has been suggested that the hydrocarbons that compose petroleum are formed during bacterial decomposition, but examination by Trask of hundreds of samples of all kinds of sediments from all over the world failed to show a single occurrence of liquid hydrocarbon. On the other hand, ZoBell and associates have shown that marine microorganisms exist which oxidize petroleum hydrocarbons.

It has been suggested that salinity is essential for accumulation of the sediments from which the liquid hydrocarbons may be produced and that salinity plays some part in changing the raw fats, waxes, gums, and resins to petroleum. Against this view is the fact that extremely bituminous sediments do accumulate in fresh water. It is probably true that high salinity has conserving properties, but on the other hand high salinity renders it difficult for organisms to live and produce deposits. Although it is true that many oil shales accumulated in, or in close proximity to, the sea, there seem to be no good reasons for believing that salinity had a great deal, if anything, to do with conservation of the organic matter or its transformation into hydrocarbons.

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CHAPTER XII

SEDIMENTS OF CHEMICAL DEPOSITION

EVAPORITES

Introduction

The term evaporites[1] is applied to a group of sedimentary deposits whose origin is largely due to evaporation. A few are included that develop through metamorphism of other evaporites. Only gypsum, anhydrite, and rock salt are of great quantitative importance.

The various more or less common evaporites are placed in five divisions based on chemical character. These, together with their chemical composition, are listed below:

Chlorides:

- Rock salt, halite, $NaCl$
- Sylvite, KCl
- Douglasite, $KFeCl_4 \cdot 2H_2O$
- Carnallite, $K \cdot MgCl_3 \cdot 6H_2O$
- Tachhydrite, $2MgCl_2 \cdot CaCl_2 \cdot 12H_2O$
- Bischofite, $MgCl_2 \cdot 6H_2O$

Sulphates:

- Alunite, $K_2O \cdot 3Al_2O_3 \cdot 4SO_3 \cdot 6H_2O$
- Alunogen, $Al_2(SO_4)_3$
- Anhydrite, $CaSO_4$
- Aphthitalite (glaserite), $(KNa)_2SO_4$
- Gypsum, $CaSO_4 \cdot 2H_2O$
- Glauberite, $CaSO_4 \cdot Na_2SO_4$
- Halotrichite, $FeSO_4 \cdot Al_2(SO_4)_3 \cdot 22H_2O$
- Polyhalite, $2CaSO_4 \cdot MgSO_4 \cdot K_2SO_4 \cdot 2H_2O$
- Krugite, $K_2SO_4 \cdot 4CaSO_4 \cdot MgSO_4 \cdot 2H_2O$
- Kieserite, $MgSO_4 \cdot H_2O$
- Epsomite, $MgSO_4 \cdot 7H_2O$
- Vanthoffite, $MgSO_4 \cdot 3Na_2SO_4$
- Blodite, (astrakanite) $MgSO_4 \cdot Na_2SO_4 \cdot 4H_2O$
- Löweite, $MgSO_4 \cdot Na_2SO_4 \cdot 2\frac{1}{2}H_2O$

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[1] Shrock has suggested that these sedimentary rocks be termed salinastones.

Langbeinite, $2MgSO_4 \cdot K_2SO_4$
Leonite, $MgSO_4 \cdot K_2SO_4 \cdot 4H_2O$
Picromerite (schönite), $MgSO_4 \cdot K_2SO_4 \cdot 6H_2O$
Kainite, $MgSO_4 \cdot KCl \cdot 3H_2O$
Celestite, $SrSO_4$
Thenardite, Na_2SO_4
Hanksite, $9Na_2SO_4 \cdot 2Na_2CO_3 \cdot KCl$
Sulphohalite, $2Na_2SO_4 \cdot NaCl \cdot NaF$

Carbonates:

Calcite, $CaCO_3$
Aragonite, $CaCO_3$
Dolomite, $CaCO_3 \cdot MgCO_3$
Thermonatrite, $Na_2CO_3 \cdot H_2O$
Natron, $Na_2CO_3 \cdot 10H_2O$
Trona, $Na_2CO_3 \cdot NaHCO_3 \cdot 2H_2O$
Gaylussite, $CaCO_3 \cdot Na_2CO_3 \cdot 5H_2O$
Pirssonite, $CaCO_3 \cdot Na_2CO_3 \cdot H_2O$
Tychite, $2MgCO_3 \cdot 2Na_2CO_3 \cdot Na_2SO_4$

Nitrates:

Soda niter, $NaNO_3$
Niter, KNO_3

Borates:

Borax, $Na_2B_4O_7 \cdot 10H_2O$
Colemanite, $Ca_2B_6O_{11} \cdot 5H_2O$
Searlesite, $Na_2O \cdot B_2O_3 \cdot 4SiO_2 \cdot 2H_2O$
Ulexite, $NaCaB_5O_9 \cdot 2H_2O$
Kernite, $Na_2B_4O_7 \cdot 4H_2O$

Pure evaporites are light colored. Rock salt, gypsum, and anhydrite form large masses. Some others occur occasionally in more or less thin beds; most are present as isolated crystals or crystal aggregates. Most evaporites result from evaporation of water of high concentration, but a few are formed by replacement, or freezing of concentrated waters. If subjected to heat and pressure, the evaporites form new combinations.

ENVIRONMENTS IN WHICH EVAPORITES ARE FORMED

Most evaporites are formed in salt lakes, arms or lagoons of the sea, and relict seas. The environments may be placed in four classes. These are (*a*) springs, (*b*) on or just beneath the surface of arid and semiarid regions, (*c*) lakes and playas, and (*d*) the sea. (Mechanical and replacement deposits are considered later.)

Evaporites Connected with Springs

Waters responsible for springs gather matter in solution from the rocks through which they circulate. Calcium carbonate seems to be the most abundant substance in solution, but in some regions it is exceeded by sodium chloride or calcium sulphate. Waters of springs have been divided into carbonate, chloride, sulphate, and other types.

Most calcium carbonate deposited about springs bears little relation to evaporation. The deposits are rarely pure and are either powdery or banded (travertine). Calcium sulphate is in somewhat the same form and is commonly known as gypsite or gypsum earth. The deposits have limited extent and cover up to a maximum of a few acres, and a maximum thickness of 4.5 meters is known. Sodium chloride due to evaporation is not common about springs.

In Alberta and Manitoba are springs that deposit sodium chloride because of freezing. Deposits of individual springs may have a thickness of about 25 cm. and may cover 50 or more square meters.

Some springs deposit alum in the form of alunite, alunogen, and halotrichite, one in Grant County, New Mexico, depositing considerable quantities of the last.

The deposits of springs are closely related in composition to that of the country rock through which the waters circulate. Thus, springs in Alberta and Manitoba derive sodium chloride from salt in the Silurian and Devonian limestones, and the springs that deposit calcium sulphate usually acquire this salt from gypsum or anhydrite deposits of the region. Related to deposits of springs are those of caves. These are chiefly composed of calcium carbonate, but many other substances are known. The deposits are in the form of stalactites, stalagmites, and kindred structures and usually are banded.

Evaporites Deposited on or Just beneath the Surface

These deposits are made in arid to semiarid regions where ground waters are brought upward by capillary action to replace water lost by evaporation. The substances in solution are derived from underlying and adjacent materials and are deposited as the waters evaporate. The deposits are known by different names in different regions: *alkali*, *caliche*, and *tepetate* (Hill and Vaughan) in different parts of the United States, *reh* in India, and *sabach* in Egypt. Some of the material known as hardpan is formed in this way. Frequently the deposits are in concretionary and pisolitic form and are firm and hard. More commonly they are unindurated and more or less powdery. They may form on or beneath the surface, and the original surface materials may become thoroughly cemented, and the deposit thus formed may be more than 1 meter thick.

The deposits consist of everything carried in solution in the rising waters, but calcium carbonate is generally the most abundant constituent of caliche. Calcium, sodium, and potassium sulphates are common, and calcium and sodium chlorides and sodium carbonate are not uncommon. Sodium carbonate is responsible for the so-called black alkali. Sodium sulphate forms the white alkali.

The most valuable salts in caliche are the nitrates, of which the two most common are those of sodium and potassium; nitrates of calcium, magnesium, and others are occasional. The most important nitrate deposits are those in northern Chile on the extremely arid lands between the coastal range and the western Andes. Rock salt and gypsum are associated and in larger quantity than the nitrates. The nitrates have their most extensive development on the western slopes of the desert area, where typically they tend to lie on lower slopes of hills rising out of the desert plain and on terraces or slopes around the salt flats (*salares*) that contain rock salt with sulphates of calcium, sodium, potassium, and magnesium. There are all gradations between the nitrates and associated salts. Figure 57 (from Whitehead) shows what are said to be typical relationships. The crude nitrate is termed *caliche* and a deposit is termed a *calcitreras*. Sodium is the most common nitrate and is ten to fifteen times as abundant as potassium nitrate. Penrose states that a descending section in the Tarapaca region is as follows:

8. Gravel with little or no salt.
7. Sandy crystalline salt, dominantly sodium sulphate.
6. Gravel cemented with sodium sulphate, termed *panqueque*.
5. Sodium sulphate containing 10 to 15 per cent sodium chloride and a trace of nitrate.
4. Sand and gravel cemented by salts containing less sulphate than 6,

but more chloride and nitrate. Zone is known as *costra* and it may be several feet thick.

3. Gravel cemented with salt of high nitrate content, the *caliche*. Thickness ranges from a few inches to several feet. Hard, compact, and with a buff to brown color.
2. Rock or gravel, uncemented, often with a high salt content and then known as *congela*, otherwise as *coba*.
1. Bedrock of stratified sands and gravels.

Zones 5 to 7 are termed *chuco* and range in thickness to several feet.

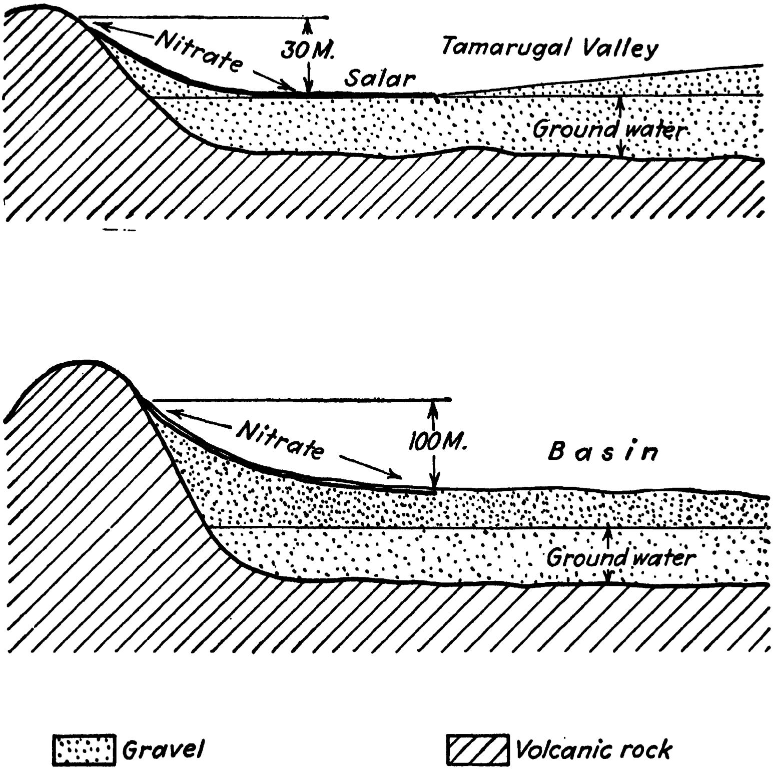

Fig. 57. Ideal diagrams showing the relations of the nitrate deposits of Chile to the topography and ground-water level. The upper diagram shows the relations in the Tarapaca region and the lower in the Aguas Blancas region. [*After Whitehead, The Chilean nitrate deposits, Econ. Geol.*, 15 (1920), 207.]

The nitrate varies from large translucent masses to interlocking crystals and efflorescences. Some deposits have consisted of 50 per cent or more of nitrate, but few such now exist.

A deposit in Bolivia has potassium nitrate the more abundant, analyses showing 60 per cent potassium nitrate and 30 per cent sodium borate. There are other deposits in South America that contain borates. Nitrate deposits in other parts of the world are limited, but all exist under climatic conditions somewhat like those of Chile.

Various explanations for the origin of the nitrogen in the Chilean nitrate deposits have been suggested, among which are derivation from guano, decomposition of algae deposited in lakes, decomposition of organic matter in soil, bacterial fixation of atmospheric nitrogen, mother liquors of salts deposited in the Andes, atmospheric electrical charges, volcanic activity, weathering of volcanic rock, and ammoniacal dust derived from the sea. However the nitrogen may have been derived, present opinion postulates that final deposition followed evaporation of ground water.

The sites of deposition are evidently due to the character of the surface and the climatic conditions. Miller and Singewald postulate that the ground waters responsible for the deposition of the nitrates originated on the western slopes of the Andes, and drained through the materials underlying the eastern part of the desert basin to its deeper western margin, where capillary action brought them to the surface and evaporation deposited the salts held in solution. The salts are assumed to have been acquired from the materials through which the waters traveled, and final deposition of the deliquescent salts was possible only over very dry areas of the desert.

Whitehead presents a somewhat different view and states that the water table in the region is so deep and the nature of the surface materials at the places of nitrate deposition of such character as to make lifting of water by capillary action impossible. Whitehead considers that the waters of deposition descended from higher levels of the slopes and that they were derived from dew, fog, and occasional rains, the quantity, though small, being adequate to dissolve the easily soluble and highly deliquescent salts. Following any deposition there would be often-repeated re-solution and downward migration of the salts, the more soluble advancing ahead of the less soluble, so that the deliquescent nitrates would be likely to precede. The first deposits made are assumed to have been mixtures of sodium nitrate, chloride, and sulphate. Cool nights with high relative humidity might lead to deliquescence of the sodium nitrate, which would then migrate downward to be deposited as the high temperature of the day lowered the relative humidity. The occasional rain might lead to solution of the sodium chloride and sulphate that had been left behind, which would then be carried downward and might be deposited above the nitrate, thus giving the three characteristic zones of the nitrate deposits, the sodium sulphate zone above, the sodium chloride and sodium sulphate zone in the middle, and the nitrate zone at the base. Whitehead favors an original volcanic source for the nitrogen. Rich considers the hypothesis of Whitehead more in accord with what is known. Singewald (1942) states that neither his (Miller and Singewald) nor Whitehead's theory is fully satisfactory and that "the facts are still too incompletely known definitely to prove or disprove either theory" and before the real explanation of origin can be attained "much more comprehensive facts must be accumulated." Rich points out that the nitrates are on surfaces that have been exposed to undisturbed weathering.

Evaporites of Lakes

The evaporites of lakes generally have composition related to the rocks of the regions in which the lakes are situated. There is hence much variation in composition. Analyses of the dissolved materials of several lakes are given in Table 26.

The Dead Sea with a salinity of 245,730 parts per million has the highest salinity of any of the large lakes.

Most evaporites of enclosed basins are due to evaporation, but some salts are precipitated by freezing of water and by changes of temperature. Freezing of water precipitates most of the salt in solution, and as changes of temperature directly affect the solvent ability of water, it results that a lowering of temperature leads to precipitation from saturated solutions. This is illustrated on the shores of Great Salt Lake and the Gulf of Karabugaz, in both of which sodium sulphate is deposited on the shores during winter months and more or less dissolved during summer months.

The evaporites of lakes may vary in character from the places of inflow of water to places where there is no inflow. Thus, calcium carbonate and calcium sulphate are precipitated where the Jordan River enters the Dead Sea, the waters of which are so highly concentrated that these salts cannot be retained in solution.

TABLE 26. COMPOSITION OF SOLIDS DISSOLVED IN LAKE WATERS

Substance	Great Salt Lake, per cent	Soda Lake, Nevada, per cent	Borax Lake, California, per cent	Lake Domoshakovo, Siberia, per cent
Cl	55.48	36.51	32.27	3.71
Br	......	......	0.04	
SO_4	6.68	10.36	0.13	63.62
CO_3	0.09	13.78	22.47	0.08
Na	33.17	36.63	38.10	30.61
K	1.66	2.01	1.52	0.59
Ca	0.16	......	0.03	0.58
Mg	2.76	0.22	0.35	0.74
P_2O_5	......	......	0.02	
R_4O_7	......	0.25	5.05	
NO_3	......	......	......	0.07
SiO_2	......	0.24	0.01	
Al_2O_3, Fe_2O_3	......	......	0.01	
Total	100.00	100.00	100.00	100.00
Salinity, parts per million	203,490	113,700	75,560	145,500

Lakes of high salinity are generally situated in dry to arid regions, but they may be present in any region where inflow of water is less than the loss due to evaporation and outflow. They are generally surrounded by areas unprotected against erosion, but surrounding areas may be covered with vegetation. During times of rainfall, lakes receive inflow of water with burdens of sediments, and these sediments become interbedded with evaporites. Lakes in arid regions may also receive considerable quantities of dust from the atmosphere. This settles to the bottom to give rise to layers of clay of history different from that brought in by water.

As lakes grow smaller through evaporation, the more insoluble salts are precipitated and gradually exposed. Exposure of precipitated salts may lead to loss of their more soluble constituents at times of occasional falls of rain. The more soluble salts are then carried to lower parts of a basin and subsequently redeposited. Thus, the less soluble salts have more extensive distribution than the more soluble.

Drifting sands may follow the retreating shores of an evaporating lake and progressively bury the salts as the waters retreat. Ultimately the entire lake area may be covered with sands.

Except under extremely rigorously arid conditions, the highly deliquescent mother-liquor salts are likely to remain in solution. If the deposits have a cover, capillary action may bring these salts toward the surface, where they may be deposited on or among the materials of the cover, or may bring the salts to the surface, whence they may be blown away by winds.

Thus, the evaporites of a desiccated lake may be expected to have an underlying and bordering deposit of the less soluble salts and a central and overlying deposit of the more soluble salts, which under most conditions would be halite, possibly more or less permeated by mother liquors.

Although it is generally true that the salts in lakes have their origin in the country rocks of the surrounding regions, there are cases where some of the salts are derived from other sources. Lake Sambhar in the Rajputana Desert of northwestern India has been proved to receive salts carried as dusts in the southwest monsoon winds, these salts having been acquired from the Cutch region to the south and west. It has been shown that at least 3,000 metric tons of salt were carried over the lake during one dry season and that 200,000 metric tons of salt are annually carried into the region of the Rajputana states.

Somewhat akin to the Lake Sambhar conditions are those postulated by Grabau to account for the salt deposited in the Silurian Salina Basin of New York. The original sources of the salts are supposed to have been in the Niagara limestones whence they were brought to the surface as caliche under arid conditions, as the limestones weathered. The caliche is assumed to have been largely sodium chloride with some calcium chloride. The sodium chloride is assumed to have been washed or blown from places of caliche formation into the deeper parts of the Salina Basin, in which the salts were deposited as the waters evaporated. Calcium sulphate is not supposed to have been in the caliche. Grabau advanced the foregoing explanation to account for the supposed occurrence of rock salt without calcium sulphate below. It has since been shown that the generally supposed absence of calcium sulphate is doubtful and certainly incorrect in some cases. Dana and later Grabau, assuming that the gypsum and anhydrite of the Salina region lie above the rock salt, postulated that formation was due to the alteration of limestone. Although it is known that local occurrences are made in this way, it is doubtful if this explanation applies to much of the gypsum in the Salina Basin. It is considered probable that the evaporites in the Salina Basin were deposited in connection with the Silurian Sea and that the region existed as a lake only in the final stages of its history.

Fossils should be generally rare in the salt deposits of desiccating lakes. Organisms would find the conditions difficult, and entrance of organisms that lived elsewhere would be accidental.

Basins of arid and semiarid regions may alternate between the conditions of playa and lake, the periods of playa and lake alternating with the weather conditions, so that for a season or a term of years a basin is a mud-cracked flat and at other seasons or terms of years it is covered with water. Thus, the Black Rock Desert of northwestern Nevada has been described by Russell as being on one morning a mud-cracked flat and in the afternoon a lake with several feet of water.

The deposits of Searle's Lake (Gale) and Death Valley, California, are those of desiccating lakes. During the Pleistocene, Searle's Lake had a depth of about 190 meters, but at present it contains water only after wet seasons, and then only to a depth of a few centimeters. At other times it is a mud-cracked playa of which the central area of 11 to 12 square miles is covered with a thick salt crust. This is surrounded by an area of approximately 50 square miles that is without vegetation and is underlain by salt-impregnated and salt-encrusted muds and sands. Alluvial fans and kindred deposits border this belt. The salt covering of the central area is mostly halite, but contains sodium sulphate, carbonate, and bicarbonate and has a thickness ranging to over 30 meters. The salt is stratified and contains more or less detrital matter. Mother liquors bathe the salts from top to bottom and are estimated to constitute about 25 per cent of the whole. This mother liquor consists of about 50 per cent sodium chloride, 20 per cent sodium sulphate, 13 per cent sodium carbonate, 12 per cent potassium chloride, 3 per cent sodium biborate, and a small quantity of sodium arsenate. Minerals identified from the Searle's Lake deposit are anhydrite, borax, celestite, colemanite, dolomite, gaylussite, glauberite, gypsum, halite, hanksite, mirabilite, natron, northupite, pirssonite, searlesite, soda niter, sulphohalite, thenardite, trona, and tychite.

Deposits of salt in lakes may range from pure to very impure; they tend to be well stratified and without fossils. The plan of a deposit is more or less circular to elliptical.

EVAPORITES CONNECTED WITH SEAS

It seems probable that most of the great salt deposits of the geologic column were in some way connected with marine waters, and it is considered that these, following Grabau, may be placed in the four classes of marginal salt pans, marine salinas, marginal lagoons, and relict seas.

Marginal Salt Pans. An artificial salt pan is a basin into which salt water is led and permitted to evaporate. Several large natural salt pans similar to those of human construction are found in the Rann of Cutch, on the Red Sea Coast, the Nile Delta, the coast of the Black Sea, and on a small scale on many coasts. That of the Rann of Cutch is the largest and is typical (Fig. 58).

The Rann of Cutch is an arid low plain south of the mouth of the Indus in northwest India. The width is about 60 miles, the length exceeds 180 miles, and the area is about 1,000 square miles. The surface is always moist, and there are occasional pools of salt water. The Rann merges into the Put, which has a dry surface, and this in turn is bordered by a sand-dune area known as the Thurr. Dunes are said to reach heights of 120 meters. Lakes may lie between them. No plants grow on the Rann and the Put, and the region contains no fresh water. The Rann is flooded to a depth of about 1 meter at the time of the southwest monsoon. Much of this water flows back into the sea when the southwest monsoon ends, and that which remains on the Rann evaporates and deposits a crust of salt of an average thickness of about 10 cm. but may exceed 1 meter. Under conditions of a stable crust it does not seem that a very great thickness of salt could be deposited in salt pans. Settling of a salt-pan area or rise of sea level may permit a thickness proportional to the extent of movement. The salts deposited would consist of calcium carbonate, calcium sulphate, and sodium chloride; mother-liquor salts are not likely, as these are washed out with each departure of water. Marine fossils may be expected. No salt-pan deposits seem to be known in the geologic column.

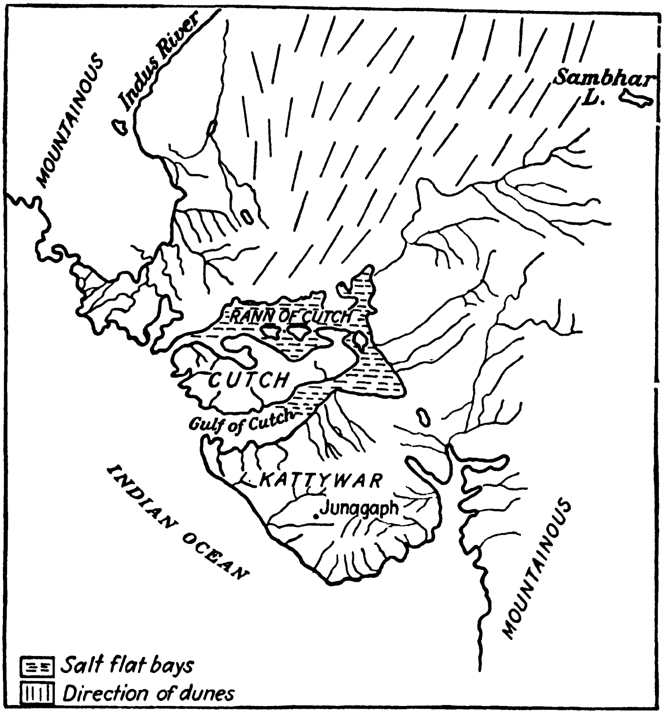

FIG. 58. Diagram showing the Rann of Cutch. Sambhar Lake is shown in the northeast corner of the diagram. The region of Kathiawar, or Kattywar, mentioned in the text as the place of occurrence of aeolian limestone is also shown. (*After A. W. Grabau, Principles of salt deposition*, 117, 1920.)

Marine Salinas. A marine salina is defined as a body of salt water adjacent to the sea and separated therefrom by a sand or gravel barrier through which it receives water from the sea. There should be little or no inflow of fresh water. Marine salinas are found on arid coasts, and some deposition of salt may be made in them. The best described example is that at Larnaca on the Island of Cyprus in the eastern Mediterranean. This is a salt lake separated from the Mediterranean by a barrier with a width of about 600 meters (Fig. 59). The barrier is composed of unconsolidated pervious materials, and water enters from the Mediterranean.

The salt lake has an area of about 2 square miles, a maximum depth of about 1 meter, and a water level about 2 meters below that of the Mediterranean. The fresh water falling in the region is largely diverted from the lake so that inflow of fresh water is exceeded by that lost through evaporation. Seepage from the sea brings in salt water, and salt is precipitated. Another example is Lake Tekir-Ghoil in Rumania. This is separated from the Black Sea by a barrier with a width of about 300 meters.

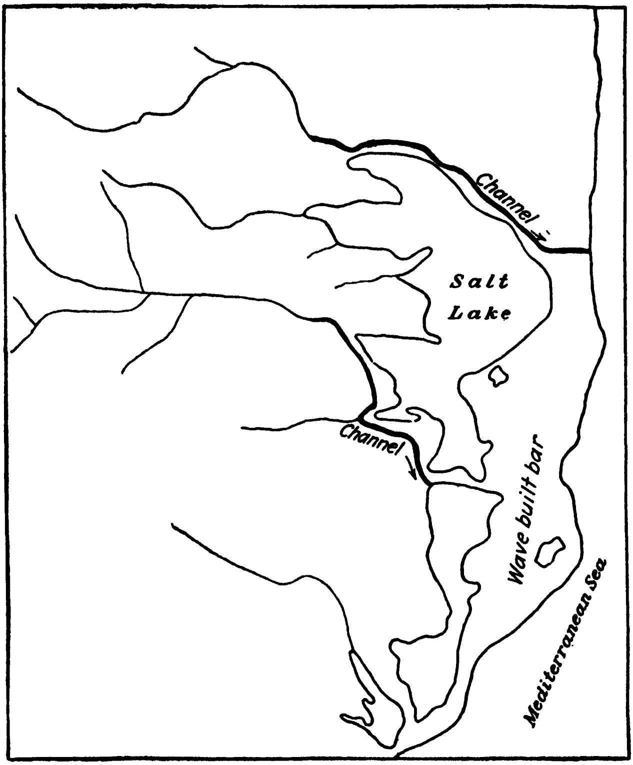

Fig. 59. Diagram of the marine salina at Larnaca on the Island of Cyprus. The channels bordering the lake are designed to drain away the fresh water derived from the region inland from the lake. [*After C. V. Bellamy, Description of the salt lake at Larnaca in the Island of Cyprus, Quart. Jour. Geol. Soc.*, 56 (1900), 39.]

Salt deposits of a marine salina are of limited area and thickness, and mother-liquor salts are unlikely. The high salinity prevents organisms from living in a salina. The sea, however, is not far away, and waves may bring organisms in time of storm, and they may occasionally be brought by animals. No deposits of marine salinas are known in the geologic column.

Lagoons. Evaporites may be deposited in the lagoons of salt lakes and the sea if conditions in the lagoons are such that the water lost by evaporation exceeds that introduced by streams and precipitation. Under such conditions currents flow into lagoons from the parent bodies, and, if evaporation is very great, the currents may be very strong. Concentration of the waters may become very high. The barrier separating a lagoon from its parent body may result from many causes, but under most conditions it is thrown up by waves. Connection with the parent body should be sufficiently large to permit adequate inflow of water, but not so large or deep as to permit circulation from the lagoon to the parent body. After the lagoon and a suitable barrier are established the constant addition of salts produces the concentration permitting deposition of calcium carbonate adjacent to the entrance into the lagoon and perhaps calcium sulphate in the more distant parts. Further concentration may lead to deposition of salts of great solubility in distant parts of the lagoon and precipitation of calcium carbonate, calcium sulphate, and some sodium chloride adjacent to the entrance. If the barrier is wave built, it may be closed at times, and the lagoon then becomes a marine salina. If there is little or no seepage through the barrier, mother-liquor salts may ultimately be deposited. Opening of the barrier initiates a new cycle of salt deposition.

The areal extent of lagoonal deposits may be large, and the thickness may equal the depth of the lagoon. A greater thickness is possible if subsidence of the lagoonal area should take place. This is not likely for a small area. Should depression take place, a thick deposit of calcium sulphate may be made in a lagoon adjacent to the entrance and a somewhat thinner deposit at the base of the lagoonal section. The calcium sulphate at the entrance should grade laterally into mingled sodium chloride and calcium sulphate, and in remoter parts of the lagoon the deposits may be dominantly sodium chloride and perhaps other salts of easy solubility.

Organisms would not be likely to live in a lagoon with highly concentrated waters, but nekton and plankton would be brought in by the entering currents in numbers related to the organisms of the parent body. These organisms would die shortly after entrance, and as there are no bottom scavengers, there would be likely to be much entombment of organic remains, and hence fossils may be rather common in lagoonal evaporites.

The lagoonal explanation for the origin of evaporites was emphasized by Ochsenius, and as example he gave the Gulf of Karabugaz, situated on the east side of the Caspian Sea (Fig. 60)

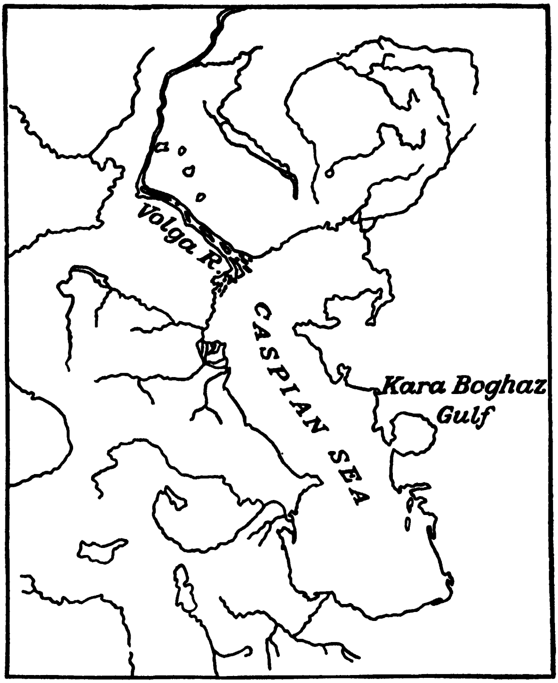

FIG. 60. Diagram showing the Caspian Sea and its drainage. The Gulf of Karabugaz (Kara Boghaz) is on the east side. It receives no inflow of fresh water and the surrounding area is very arid.

This sea is surrounded by a more or less arid region, and the aridity is particularly rigorous on the east side. The Caspian is a relict sea that at the time of separation from the parent body evidently had the same degree of salinity. Since the separation it has received additions of salts from the entering streams, and there has been much evaporation, as the surface is below sea level. Nevertheless, its salinity is less than that of the Mediterranean and the ocean.

The explanation for the low salinity of the Caspian is found in the Gulf of Karabugaz. This body of water, an appendage of the Caspian Sea, is about 95 miles long, 80 miles wide, and not over 15 meters deep. It is separated from the Caspian by a barrier consisting of two sandspits. Its salinity is 163,960 parts per million as compared to the Caspian with 12,670 parts per million. Evaporation in the gulf is so great that a strong current constantly flows into it from the Caspian. This current has built a delta at the entrance and daily brings in 350,000 tons of salt. The waters of the gulf are estimated to contain 34 billion million metric tons of salt.

At the present time calcium sulphate (gypsum) is deposited in large quantities adjacent to and for some distance beyond the opening into the gulf. The waters are saturated with sodium sulphate during the winter season, and that salt (glauberite) is then deposited on the shores and shallow bottoms over an estimated area of 3,500 square km. It is partly dissolved during the summer. The concentration necessary for the deposition of sodium chloride has not been reached.

The current flowing into the gulf teems with aquatic organisms of which most die shortly after entrance. The skeletons of many of these and also of the birds that feed upon them cannot fail to be deposited with the salt, and the deposits of this lagoon may be expected to be abundantly fossiliferous, particularly adjacent to the place of entrance from the parent body.

No lagoonal evaporites are known in the older parts of the geologic column. The construction of the Suez Canal through the comparatively recent deposits of the great Bitter Lakes of Suez exposed a deposit of salt over an area 8 miles long and 4 miles wide. These lakes represent a former lagoon. The salt in the central part is 60 ft. thick and composed of layers of halite and gypsum separated by beds of clay containing shells of organisms of species still living in the Red Sea. The salt layers range in thickness from 3 to 18 cm., and 42 layers of halite and gypsum were counted in a thickness of 2.46 meters. The repeated occurrence of layers of fossiliferous clay proves repeated entrances of the sea. Mother liquor was present in the salts, but in far less quantity than was demanded by the salts that had been deposited.

A Miocene salt deposit in Wieliczka, Galicia, is abundantly fossiliferous and carries no mother-liquor salts. It may be considered the oldest known example of a lagoonal deposit.

Branson assumed that several connected basins would be necessary to explain the salt deposits as he understood them. His theory rests on the fact that some evaporite deposits do not have the sequence of salts that should exist if ocean waters evaporated and that most do not contain fossils. He assumed that the waters flowed from a parent source into the first of several succeeding basins and with increasing concentration moved into others. The organisms were killed in the first basin, and the calcium and magnesium carbonates and perhaps some of the calcium sulphate were deposited in the first basin. The concentrated waters then flowed into a second basin and progressively into others, depositing salts related to concentration in each. This explanation requires an extremely nice balance of conditions that probably did not often occur, and it is considered very unlikely that the extensive deposits of evaporites developed in this way.

Relict Seas. Relict seas are bodies of water that have become separated from the parent body by diastrophic, depositional, or volcanic causes. Immediately following separation from the parent body the waters of a relict sea may approximate or exceed the salinity of the parent body, but gradually the salinity changes, the waters becoming less saline if the sea is fed by streams so as to have outward drainage, and more saline and of lower level and reduced area if the surrounding region is semiarid to arid. It is a relict sea under either condition, but only in the latter case is there any possibility of deposition of evaporites. Well-known relict seas are the Salton Sea of California with a depositional barrier, the Caspian Sea, Lake Nicaragua, and Lake Baikal.

The Caspian Sea once covered a much larger area, had a higher level, and was reduced to its present level of 25.6 meters below sea level through evaporation. This did not result in greater salinity because of movement of the salt into the Gulf of Karabugaz and inflow of fresh waters from the Volga and other rivers. The abandoned part of the former bottom has over 2,500 lakes and playas, of which most are salty with a considerable range in composition. During wet seasons the more soluble salts over abandoned parts of the basin are washed into the lakes and playas or into the Caspian. If there was no Gulf of Karabugaz, if the surrounding region was so dry that there was little or no inflow of fresh water, and if there was a narrow connection with the Black Sea of such restriction as to prevent circulation, sea level would be maintained and the salinity would become very great. This would ultimately result in precipitation of calcium carbonate near the entrance, calcium sulphate farther inward, sodium chloride over remote parts of the basin, and perhaps bitter salts in remote and restricted bays. Calcium carbonate followed by calcium sulphate ordinarily would underlie any sodium chloride, but the greater thicknesses of the carbonate and the sulphate should be adjacent to the entrance.

Separation of a desiccating relict sea from the parent body would produce a body of water that from its beginning had high salinity, and this would increase as evaporation lowered the surface and the occasional rains washed salts from abandoned parts of the basin. Following deposition of sodium chloride would come deposition of the more soluble salts. Continued evaporation would reduce the relict sea to a playa or series of playas in which the mother-liquor salts would be concentrated either to be deposited or to impregnate the salts already deposited.

As relict seas may cover large areas, they may undergo some changes of position with diastrophism and from time to time invade areas not previously covered, so that salts of wide range of composition may be the basal or initial deposits of a section deposited in a relict sea. Diastrophism may also deepen a basin before connection with the parent sea is severed, and thus a thickness of salt related to the original depth and the extent of deepening may be deposited. Deep relict seas exhibit density stratification, as shown by the Dead Sea in which the salinity at the surface is 19 per cent or 192,150 parts per million and 26 per cent or 259,980 parts per million at the depth of 300 meters. This might lead to deposition of different salts on bottoms of different depths.

Deposits of desiccating relict seas may be fossiliferous as long as salinities not prohibitive to organisms existed, and fossils may be abundant in deposits made adjacent to entrances, for as long as connection with the parent body obtained.

Rainfall; temperature; duration of connection with the parent body; shape, depth, and size of the relict sea; inflow of clastic sediments; and diastrophism are factors that lead to variations in the character, thickness, and succession of evaporites; and, as there are at least eight important variables, it follows that there are many possible variations in sequence and gradations of deposits.

It is the considered opinion that the great deposits of evaporites in the geologic column were deposited in relict seas. The general features of these evaporites and the associated sediments indicate that deposition did not take place in very deep water, but, as the sequences of sediments are hundreds and even thousands of feet in thickness, it follows that the basins subsided as the sediments accumulated.

The sea in which the evaporites and associated strata of the Permian basin of Kansas and Oklahoma accumulated was once in open connection with the waters to the southwest, but ultimately it seems that the opening or openings were restricted and finally closed by the mountains of southern Oklahoma, northern Texas, and southeastern Colorado and by deltas built by streams flowing from these mountains. The surrounding region was dry. It is known that calcium sulphate was deposited over parts of this basin before the deposition of sodium chloride and also contemporaneous and subsequent thereto, but with possibly minor exceptions the concentration necessary for deposition of the mother-liquor salts does not seem to have been attained. In some localities sodium chloride has been stated to form the basal evaporites, explained by Haworth as due to warping of the basin and overflowing of areas not previously submerged.

The Permian basin of west Texas and New Mexico is also considered to have been a relict sea that had connection with the open ocean to the southwest and was separated from the Atlantic by a highland area in east and southeast Texas. This basin contained more or less normal marine waters in the early part of the Permian. The surrounding regions seem to have been of some degree of aridity, so that evaporation exceeded additions of water from streams and precipitation, but the water level did not change as long as there was connection with the open ocean to the southwest. Algal reef limestones formed over elevated areas on the bottom, divided the bottom into different basins, impeded circulation, and limited inflow from the ocean. Evaporite deposition began in the eastern basin and progressed westward. Ultimately a high degree of salinity was attained, connection with the ocean was severed, and there was deposition of sodium chloride and later of polyhalite and other potash minerals.

A very interesting gypsum in the Permian of this area which is unique in containing euhedral crystals of quartz was brought to the writer's attention by Professor A. N. Murray of the University of Tulsa. Specimens of the gypsum with the quartz phenocrysts were collected by a student from a locality 5 miles east of the town of Lake Arthur, Chaves County, New Mexico, in the outcrop area of the Chupadero formation (Permian). The doubly terminated quartz crystals described by Tarr may have come from the same formation as they seem to have been collected in the same general area. Professor Murray's description of the specimens is as follows:

"These quartz crystals are embedded in a matrix of white, finely crystalline gypsum. A thin zone of gypsum around the quartz is more coarsely crystalline than the matrix. Usually the clinopinacoidal cleavage of the gypsum grains in this zone can be readily seen with the unaided eye.

"The crystals of quartz occur as aggregates and as euhedrons scattered through the gypsum like plums in a pudding, . . . distance between individual aggregates and grains varies from 0.6 to 1.6 inches. The color of the quartz is smoky gray with occasional reddish tinges. The aggregates are roughly tabular in shape with the large dimension varying between 0.4 and 1.0 inch. The individual crystals are usually distorted euhedrons that are doubly terminated. They may be as much as an inch in length parallel to the "c" axis and about half as thick as they are long. There is a tendency for the crystals to be flattened to one of the prism faces." There seems to be no relation between the flattening of the aggregates and euhedrons to the bedding.[1]

Tarr considered the crystals described by him to have been replacements of gypsum by silica derived from associated sediments and deposited from solution.

The Germanic Permian basin of north central Europe is also interpreted as a relict sea that was produced by elevation of the Hercynian Mountains across central Europe. These separated the Germanic basin from the Atlantic on the west and the Mediterranean on the south. Connection with the parent waters is thought to have been toward the east. The region was arid, and evaporation was in excess of additions of fresh water. This permitted inflow from the sea and ultimately produced high salinity.

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[1] Professor A. N. Murray, letter of May 1, 1948.

Closing of connection with the sea finally led to complete evaporation of the waters and deposition of mother-liquor salts.

MECHANICAL AND ALTERATION DEPOSITS OF EVAPORITES

So far as is known, mechanical deposits of evaporites are composed of calcium sulphate and consist of particles of gypsum or anhydrite of sand dimension. These were derived from older gypsum or anhydrite deposits or from efflorescences formed on the surface or about springs and lakes. The deposits are in the form of dunes and occurrences are few. The most important is in Otero County, New Mexico, where dunes of gypsum are distributed over an area of about 300 square miles (Herrick, Talmadge). Another small occurrence is in Utah in which the particles are small crystals of gypsum with well-preserved crystal faces (Gilbert). Salts of similar mechanical deposition, but on a smaller scale, are said to be present in Australia. It was suggested by Wilder that some of the ancient deposits of calcium sulphate may have developed in this way, and he called attention to cross-laminated gypsum at one locality in Oklahoma.

It is considered that mechanical deposits of evaporites are of limited quantitative importance, and that very few, if any, deposits of evaporites in the geologic column developed in this way.

Alteration deposits of salts among those listed as evaporites are largely calcium sulphate, although it seems certain that some of the salts found in the Stassfurt region of Germany are salts that developed from recombinations of those originally present. Sulphuric acid is formed on oxidation and hydration of sulphides, and if this flows over carbonates, these react with the acid, and sulphates result. This is known to have taken place to a considerable extent in some of the gypsum areas of New York, and Dana and Grabau explained the gypsum deposits of that state in this way. Wilder also held replacement responsible for some gypsum deposits. It is not considered likely that many extensive gypsum deposits had this origin.

SALTS OF ORGANIC PRECIPITATION

Most plants require potash, which they take from the soil in which they grow and return to the soil when they decay. The ash of corncobs has been found to contain a maximum of 20.13 per cent of water-soluble potash and 4.01 per cent phosphoric acid, the ash of gooseberry canes 13 per cent potash, and yellow banana stalks 49.40 per cent. The ash of kelps may contain 20 per cent potash and over 1.5 per cent iodine. The ash of *Salsola kola* L. of the Spanish Coast may contain 14 to 20 per cent sodium carbonate. Similar soda-using plants grow in the beaches of the Canary Islands, the Argentine salt plains, and other parts of the world where sodium is common in the soil. Some plants secrete sodium chloride, and in some parts of the world the ash of these plants is said to be leached to obtain salt.

These occurrences are interesting, but the salts used by plants cannot be expected to form a deposit of salt, although they may indirectly aid in doing so. Potash deposits of some Nebraska lakes have been ascribed to leaching of ash produced by burning of the prairie vegetation of the areas surrounding the lakes.

EVAPORATION OF SEA WATER

Evaporites are assumed mostly to have formed from evaporation of sea water. The composition of sea water is therefore of considerable importance, and it may be assumed that whatever takes place when sea water evaporates is also applicable to lake water, with modification due to differences of composition.

Except for variations related to proximity to mouths of great rivers and areas of great precipitation or evaporation, the average salinity of the ocean is about 35,000 parts per million, or about 3.5 per cent by weight. The salinity of the Dead Sea is much greater, 192,150 parts per million at the surface and 259,980 parts per million at the depth of 300 meters. Great Salt Lake has a salinity ranging from 137,900 to 277,200 parts per million. The elements in solution in sea water are given in Table 27.

Other elements detected in sea water are aluminum, barium, boron, cesium, cobalt, copper, iron, lead, lithium, manganese, nickel, radon, rubidium, strontium, and zinc. Sodium and chlorine compose nearly 91 per cent of the total. There is more than three times as much magnesium as calcium, more than twice as much sulphur as calcium, and the potassium in solution equals that of calcium. According to Moberg, the elements in the lower part of the table bear a constant ratio to each other and the total salt. Possible combinations of the elements in solutions are shown in Table 28. The small quantity of calcium carbonate should be noted.

As sea water evaporates, the salts are precipitated in reverse order to solubility with such modification as may be imposed by temperature, sunlight, relations to each other, and perhaps other factors.

Usiglio evaporated sea water obtained in the Mediterranean off the south coast of France, 5 liters being used in each experiment, one sample being obtained at 1-meter depth 3,000 meters from the shore and another at the same depth 5,000 meters from the shore. The initial salinity was 38,450 parts per million. The waters were evaporated in large porcelain dishes in a room maintained at 40°C. and the air kept dry by means of quicklime. From time to time the liquids were decanted and the precipitates dried and weighed. The results obtained are given in Table 29.

TABLE 27. ELEMENTS IN SOLUTION IN SEA WATER

Element	Parts Per Million
Cl	19,400.00
Na	11,400.00
Mg	1,400.00
S	900.00
Ca	400.00
K	400.00
Br	70.00
C	30.00
F	0.80
Si	0.20
N	0.25
I	0.05
As	0.01
P	0.02
Ag	0.01
Au	0.000005

TABLE 28. COMPOUNDS IN SOLUTION IN SEA WATER

Compounds	Parts per Million
$NaCl$	27,213
$MgCl_2$	3,807
$MgSO_4$	1,658
$CaSO_4$	1,260
K_2SO_4	863
$CaCO_3$*	123
$MgBr_2$	76

* Includes all traces of other salts.

As the last bitterns were evaporated it was found that temperature has significant effects. Cooling accelerated magnesium sulphate precipitation, but a part of this reentered solution if the waters were warmed. Continued evaporation precipitated sodium chloride and magnesium sulphate, and cooling led to precipitation of more magnesium sulphate. The differences in temperature between day and night produced interesting effects as the last bitterns were evaporated:

Table 29. Order of Precipitation of Evaporites in Usiglio's Experiments

Volume	Weight, grams							
	Fe_2O_3	$CaCO_3$	$CaSO_4 \cdot 2H_2O$	NaCl	$MgSO_4$	$MgCl_2$	NaBr	KCl
1.000								
0.533	0.0030	0.0642						
0.316	......	Trace						
0.245	......	Trace						
0.190	......	0.0530	0.5600					
0.1445	......	......	0.5620					
0.131	......	......	0.1840					
0.112	......	......	0.1600					
0.095	......	......	0.0508	3.2614	0.0040	0.0078		
0.064	......	......	0.1476	9.6500	0.0130	0.0356		
0.039	......	......	0.0700	7.8960	0.0262	0.0434	0.0728	
0.0302	......	....	0.0144	2.6240	0.0174	0.0150	0.0358	
0.023	......	......	......	2.2720	0.0254	0.0240	0.0518	
0.0162	......	......	......	1.4040	0.5382	0.0274	0.0620	
Total.......	0.0030	0.1172	1.7488	27.1074	0.6242	0.1532	0.2224	
Salts in last bittern...	......	......	......	2.5885	1.8545	3.1640	0.3300	0.5339
Total solids.	0.0030	0.1172	1.7488	29.6959	2.4787	3.3172	0.5524	0.5339

1. Night: epsomite.
2. Day: hexahydrate, halite, rarely potash salts.
3. Night: epsomite.
4. Day: hexahydrate, picromerite, halite, bischofite, magnesium bromide
5. Afternoon: carnallite.
6. Night: carnallite, picromerite.
7. Day: little precipitation.
8. Afternoon: carnallite.
9. Night: carnallite, epsomite.
10. Autumn: several months later, temperature 5 to 6°C., bischofite.

Usiglio's results showed that salts are precipitated from sea water in the following order: (*a*) calcium carbonate and ferric oxide; (*b*) calcium carbonate overlapping calcium sulphate; (*c*) calcium sulphate overlapping calcium carbonate in the beginning and sodium chloride at the end; and (*d*) sodium chloride more or less coinciding with the precipitation of magnesium sulphate and chloride and sodium bromide with precipitation continuing into the precipitation of the salts of the last bitterns.

Based on Usiglio's experiments the sequence of deposits of evaporites should be calcium carbonate at the base, followed in succession by calcium sulphate, sodium chloride, and lastly by sodium chloride with the salts of last bitterns. This succession does not always obtain. There are deposits of calcium sulphate without calcium carbonate at the base, deposits of sodium chloride without underlying calcium sulphate. These conditions may be explained as caused by invasion of concentrated waters over areas not previously submerged.

Usiglio's experiments failed to obtain several minerals found in deposits of evaporites. Among these are kieserite, polyhalite, kainite, and anhydrite. This may be explained as a consequence of the short time involved and the relatively low temperature. He also failed to obtain pure sodium chloride and found that pure calcium sulphate was precipitated over a very narrow range, which is in harmony with the statement of Wilder and Stieglitz that all calcium sulphate deposits formed under existing conditions of the atmosphere must contain 0.9 per cent of calcium carbonate.

It is probable that under natural conditions precipitation of evaporites has rarely been brought to completion, that in some bodies precipitation did not extend beyond the carbonate stage, in others it did not extend beyond the calcium sulphate stage, in others it proceeded to the sodium chloride stage, and there must have been few cases where there was complete evaporation of a large body of sea water to the extent that the mother-liquor salts were deposited. Furthermore, it is difficult to understand how they could be preserved after precipitation, as only a small quantity of moisture is required to return them into solution. Extremely low relative humidity and quick burial beneath impervious sediments seem essential.

The two forms of calcium sulphate, gypsum and anhydrite, require different conditions for formation. Van't Hoff and

Weigert experiments showed that precipitation of anhydrite and gypsum was a function of temperature and presence of other salts in the evaporating solution. Below 66°C. the precipitate was gypsum; above that temperature it was anhydrite. Thus, in a solution containing calcium sulphate alone, the salt precipitated under natural conditions would probably be gypsum. Other salts in the solution change the critical temperature, and such would be the case under most natural conditions. If sodium chloride was in the solution, the critical temperature would fall to 30°C., and if the sodium chloride was concentrated, the critical temperature would fall to 25°C. The work of Posnjak showed that the temperature of 66°C. is not correct and that the critical temperature is 42°C. Sea water is not saturated with calcium sulphate, and concentration is necessary before calcium sulphate salts in any form are precipitated. The solubility of both gypsum and anhydrite increases to about the normal salinity of sea water, and with further increase in concentration there is decrease in solubility with the solubility of anhydrite decreasing more rapidly than gypsum. Posnjak's experiments, conducted with artificial sea water containing the essential salts, at 30°C. showed that the salt content must be increased 3.35 times the usual salinity before precipitation of calcium sulphate begins and only about one-half of this salt is precipitated as gypsum up to the concentration of 4.8 times the usual salinity when precipitation of anhydrite begins. As much evaporation of normal sea water is below the temperature of 30°C. (86°F.), it would seem that considerable calcium sulphate in the geologic column was originally deposited as gypsum. However, as most evaporites were deposited under arid and semiarid conditions with temperatures fairly high and with a high content of sodium chloride in the evaporating waters, it is probable that the greater part of the calcium sulphate in the column originally was in the form of anhydrite.

If anhydrite is exposed to atmospheric conditions, it changes to gypsum, and if gypsum is subjected to heat and pressure, it changes to anhydrite.

The most intensive and extensive studies of the evaporites were made by Van't Hoff and his associates, and evaporation of solutions containing single and two or more salts and the effects of temperature and other factors were studied.

It was found that, at 20°C., 100 grams of water dissolves 26.4

grams of sodium chloride or 25.6 grams of potassium chloride, and, at 50°C., 26.8 grams of sodium chloride or 30 grams of potassium chloride. With the two salts in solution at 20°C., 100 grams of water dissolves 20.3 grams of sodium chloride and 10.2 grams of potassium chloride, and, at 50°C., only 18.5 grams of sodium chloride, but 14.7 grams of potassium chloride. If the solution of the two salts at 20°C. was raised to the higher temperature, there would be precipitation of sodium chloride. In a complex solution like sea water there are probably different degrees of solubility for each substance at different temperatures, and any marked change of temperature would lead to precipitation of some substances from a saturated solution. Table 30 lists temperatures of formation of several double salts in the Stassfurt deposits and the solutions from which they were precipitated.

TABLE 30. TEMPERATURES OF FORMATION OF SEVERAL SALTS

Salts	Molecules in 1,000 molecules H_2O						
	$NaCl$	KCl	$MgCl_2$	Na_2SO_4	K_2SO_4	$MgSO_4$	Temperature, °C.
Carnallite	..	6	89	..	..	..	25
Schönite	..	..	..	..	11	40	25
Glaserite	88	30	..	9	..	..	25
Kainite	11	12	61	..	..	12	25
Leonite	24	20	40	..	..	17	25
Astrakanite	75	..	2	..	..	27	25
Langbeinite	43	42	36	..	..	11	83
Loewite	53	..	34	..	..	12	83
Vanthoffite	86	..	13	..	..	12	83

It should be noted that langbeinite, loewite, and vanthoffite form at a temperature that is much higher than is normal in surface waters, and it is probable that these three minerals formed subsequent to deposition and burial.

A solution of sodium chloride, potassium chloride, magnesium chloride, and magnesium sulphate in the proportion of 24:15.5:40.7:20 in molecules per thousand molecules of water precipitated on evaporation the following salts together with sodium chloride in the order named:

1. Magnesium chloride.
2. Magnesium sulphate and kainite.
3. Hexahydrate and kainite.
4. Kieserite and kainite.
5. Kieserite and carnallite.
6. Kieserite, carnallite, and magnesium chloride.

Normal sea water at 25°C. which contained, in molecules per thousand molecules of water, 100 sodium chloride, 2.2 potassium chloride, 7.8 magnesium chloride, and 3.8 magnesium sulphate, and saturated for sodium chloride but not for the other salts on evaporation precipitated in sequence the salts shown in Table 31. It will be noticed that, whereas all sodium chloride was precipitated, 2.2 molecules of the other substances remained in solution.

TABLE 31. PRECIPITATION OF SALTS ON EVAPORATION OF SEA WATER AT 25°C.

Sequence	Halite	Kieserite	Kainite	Carnallite	Bischofite
1	95.40				
2	4.42	1.05	2.02		
3	0.03	0.35	....	0.10	
4	0.15	0.38	....	0.08	7.62
Totals.........	100.00	1.78	2.02	0.18	7.62

Sea water has potassium and magnesium in the ratio of 1 to 10.9, and the sulphate radical to magnesium as 1 to 3. These ratios do not obtain in salt deposits and are respectively 1 to 4.33 and 1 to 1.57. The explanation may be found in the last bitterns, which are commonly not completely deposited.

If the different salts present in sea water were all deposited, they would be in the proportions shown in the first part of Table 32, but these proportions should not be expected in salt deposits, as there are many things that change the proportions. The proportions that exist in the Stassfurt Region are shown in the last half of the table. The increase in calcium sulphate and decrease in magnesium and potassium are probably connected with solubilities of the salts. Sodium chloride is taken in each case as 100 meters.

Complete evaporation of a 300-meter column of sea water would make a deposit of about 4.5 meters of salt of which 2.1 cm.

would be calcium sulphate. To form a bed of calcium sulphate 3 meters thick, a very common thickness, requires evaporation of a column of sea water about 4,200 meters high, and to produce the more than 360 meters of anhydrite found in a Texas oil well would require a column over 300,000 meters high. No bodies of water so deep were ever evaporated; and the great thicknesses of calcium sulphate and sodium chloride in the column are due to development of high salinity in relict seas by inflow from parent bodies before connections were severed; and subsequently as the waters were evaporated and the bottoms were uncovered, salts previously deposited were washed into the central parts of the basin, so that the places of deposit of the evaporites contain the salts that were in a body of water that once covered a much larger area.

TABLE 32. THICKNESSES OF SALTS FORMED FROM EVAPORATION OF SEA WATER AND THE ACTUAL THICKNESSES IN THE STASSFURT REGION

Salt	Thicknesses precipitated from sea water, meters	Thicknesses in Stassfurt region, meters
Anhydrite	3.4	5.7*
Rock salt	100.0	100.0
Kieserite	7.2	2.2
Carnallite	14.0	4.7
Bischofite	4.7	

* This figure becomes 20.4 if all the anhydrite in the section is considered.

IMPORTANT DEPOSITS OF EVAPORITES

Thick deposits of anhydrite and gypsum are in Mississippian strata of New Brunswick, Nova Scotia, and Newfoundland, and a thickness of 75 meters is present at Hillsborough, New Brunswick. There are units of calcium sulphate in the Silurian of the Salina Basin of New York with a thickness of more than 20 meters. Rock salt in the same section has a thickness of 90 meters. Maximum thickness of beds of anhydrite and gypsum in the Permian of Kansas and Oklahoma ranges to over 18 meters. There is much rock salt in the Salina Basin of Kansas, and the maximum aggregate thickness is over 30 meters. There is much anhydrite but no halite in the Triassic Chugwater formation of Montana and Wyoming. A well drilled in the Permian of West

Texas showed 348 meters of nearly pure anhydrite, and this salt was present through 585 meters. The anhydrite beds are laminated in units that average less than 2 mm. thick. There is also much rock salt and some mother-liquor salts in the West Texas area. The Eocene of the Paris Basin contains a 19.5-meter bed of gypsum that is unusual in that it contains bones of mammals, shells, and wood. There is no halite. Evaporites are also present in the Tertiary of the Aragon Basin of Spain. Great thicknesses of evaporites are in the Permian of the Stassfurt Region on the Rhine.

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CHAPTER XIII

SEDIMENTS OF CHEMICAL DEPOSITION

PHOSPHATIC SEDIMENTS

Introduction

The original source of the phosphorus that enters sedimentary rocks is largely found in apatite, a mineral of wide distribution in igneous rocks. A little is derived from monazite and xenotime. Transportation is probably as phosphoric acid and as calcium phosphate, the latter soluble in water containing carbon dioxide and swamp water rich in organic matter. Some calcium phosphate may be transported in colloidal form.

The phosphorus is precipitated as some form of hydrous calcium phosphate. Precipitation seems to be largely done through organic agencies: either directly, in which case the phosphorus enters the bones and teeth of vertebrates and the shells of invertebrates; or indirectly through reaction of phosphorus in solution with products of organic decomposition. As the tissues of some bacteria contain phosphorus it may be that some is thus precipitated. Calcium phosphate may also be precipitated inorganically in the presence of calcium carbonate.

The two most common phosphate minerals in sedimentary deposits are collophanite ($Ca_3P_2O_8 \cdot H_2O$) and dahllite ($Ca_6(PO_4)_4 \cdot CaCO_3 \cdot \frac{1}{2}H_2O$). Deposits are either primary or secondary. Primary deposits include the stratified marine phosphorites, phosphatic nodules, and primary guano. Secondary deposits include those enriched by leaching of rocks containing small quantities of phosphate, by deposition of phosphates in other rocks, and detrital phosphates.

Primary Phosphates

STRATIFIED MARINE PHOSPHORITES

The composition of pure phosphorites approaches apatite, but no crystalline form is shown. Instead the phosphorites have powdery, oölitic, or concretionary forms. Deposits extend over thousands of square miles in formations of more or less uniform thickness. The deposits are never entirely composed of phosphate, but consist of phosphates interbedded with claystones, siltstones, and, more rarely, limestones and sandstones. The Permian Phosphoria formation of the northern Rocky Mountain Region of the United States is typical. This formation in Idaho shows 17 units of phosphate in a little less than 8 meters of section with range in thickness from 2.5 to 29 cm. Interbedded strata are mostly shales in which some phosphate is present. Underlying strata consist of more or less phosphatic shale with some limestone, and near the base is a very pure bed of phosphate with thickness of 5 meters. Maximum thickness of units in this formation containing good phosphate is 9 meters. The range in thickness of individual beds of phosphate is from about 1.5 mm. to 20 cm. with most beds less than 2.5 cm. thick. Bedding is regular and even and devoid of marks of wave and current action. Contacts between phosphatic units and the interbedded strata in some cases are sharp and in others gradational. There are streaks of shale and sandstone in phosphatic beds and streaks of phosphate in shale and sandstone. Some beds contain pebbles derived from previously deposited phosphates. Colors are gray, brown, or black, depending upon the quantity of contained organic matter. There is a more or less strong bituminous odor, and textures range from compact to powdery, oölitic, or concretionary. The concretions or nodules range to about 5 cm. in diameter and have concentric structure in whole or in part. The oölites range from mere specks to 2 mm. in diameter and in shape from spherical to irregularly flattened. Internal structure is concentric in many, but some have no concentric structure and others have it only on the periphery. Particles without concentric structure consist of more or less granular or amorphous materials, of which the exact nature has not been determined. Blackwelder suggested that these parts may be of excremental origin. The phosphatic beds contain few fossils. Only a few discinoid brachiopods, gastropods, and fragments of fish bones have been found.

No bed is composed of pure phosphate, but the content of many is very high and ranges to 35 per cent or more. Impurities consist of particles of clay and silt size of which many are composed of quartz. Marcasite and pyrite are common. Schaller states that the oölites are composed of amorphous materials which may be collophanite and that some are coated with crystallized mineral matter that may be one of several phosphatic minerals.

The matrix of the oölites is commonly of similar materials, but in some cases it consists of calcite, dolomite, and, rarely, chert.

The origin of the stratified phosphorites has not been solved, but certain facts bearing on origin stand out very clearly. Formation in a marine environment is proved by the presence of marine fossils and interbedding with marine shale, sandstone, and limestone. The marcasite and the bituminous matter prove formation under reducing conditions. Replacement of shells and bones of fish, probably originally slightly phosphatic, proves diagenesis for these constituents, and diagenesis may be expected to have affected other constituents. Depth is not clearly indicated, but there is nothing suggesting deep water except the absence of cross-lamination and ripple marks in the sandstones. If the spherules without oölitic structure and the central parts of others are excremental, it proves that at times of slow accumulation of other sediments there was extensive accumulation of excremental particles not masked by other sediments. The responsible organisms were either plankton or nekton, or both, that lived in overlying waters, or they were benthos without hard parts. The shelled benthonic population was few in kinds and numbers.

A part of the phosphorus of the phosphorites may have been derived from shells and bones of fish, but, as these remains are not common, it does not seem possible to relate much phosphorus to this source. The phosphorus seemingly must have been derived in large part directly from sea water. Murray postulated sudden destruction of vast numbers of organisms, the decay of which produced ammoniacal solutions that led to precipitation of calcium phosphate. Blackwelder also appealed to wholesale destruction of organisms and to ammoniacal solutions that dissolved phosphatic matter from bones, shells, etc. This was later precipitated as hydrous calcium carbophosphate, which replaced carbonates, increased the phosphatic content of organic matter, and formed the oölites, nodules, and the paste that now cements the oölites. He also suggested an excremental origin for the spherical particles. Breger suggested extraction of the phosphorus from sea water by bacteria and ascribed the oölites and spherulites to rolling or foraminifera. Pardee considered failure of decomposition to be due to low temperature, which also permitted carbon dioxide accumulation. This dissolved carbonates and concentrated phosphates. Mansfield suggested that the phosphatic oölites and spherulites were formed on the bottom directly from precipitation of phosphatic solutions or colloids. These may have been due to some wholesale destruction of organisms, but it seems more reasonable to assume that they represent slow accumulation of phosphatic debris under conditions that largely excluded oxygen from the deeper waters. This also eliminated benthonic organisms and scavengers other than anaerobic bacteria, which in turn permitted accumulation of organic debris. Temperatures were cool, which favored growth of plants and animals in shallower parts of the basin and reduced activities of dentrifying bacteria, which may have curtailed precipitation of calcium carbonate, which in turn favored concentration of the phosphatic solutions that led to the formation of the oölites.

Mansfield's latest contribution (1940) to the origin of the stratified marine phosphorites suggests that, since fluorine is an essential constituent of phosphatic deposits and tends to keep the phosphatic compound relatively insoluble, it probably plays some part in phosphate deposition. Since fluorine is derived largely from volcanism, deposition of phosphate minerals containing fluorine would take place under favorable conditions of environment at times of extensive volcanism and, as uncomformities bear some relation to volcanism, it is suggested that there may be some connection between uncomformities and phosphate deposits. The connection seems remote to the present writer.

Dietz, Emery, and Shepard have described phosphatic deposits on the sea bottom off the coast of southern California. These vary from thin slabs to large masses, and the deposits in any individual area have the same general characters. Colors range from dark brown to black, and surfaces are smooth and shiny. Collaphanite is the chief mineral. Detrital matter and a little organic matter are enclosed in the phosphate bodies, and the organic matter is usually replaced by francolite (a variety of calcium phosphate). X ray gave an apatite pattern. The deposits are on the sides of banks, steep escarpments, walls of submarine canyons, and on the continental shelf, all places of essentially nondepositional environments. The authors consider it probable that sea water deeper than a few hundred meters is essentially saturated with tricalcium phosphate and that there may be an annual deposition of phosphate essentially equal to that annually contributed to an area by streams. Some precipitation may have been made from colloidal suspension. The deposits were formed in place, they are in places where clastic sediments are not accumulating, and they are not erosion remnants.

PHOSPHATIC NODULES

Phosphatic nodules have considerable distribution over the present ocean bottom, where, according to Murray and Hjort, they are found along coasts subject to rapid and considerable changes of temperature. The usual depth range seems to be from 200 to 500 fathoms, but they have been dredged from depths of 1,500 fathoms.

Phosphatic nodules have extensive distribution in the geologic column, and in greater or less abundance they have been found in every geologic system beginning with the Cambrian. They are common in dark clays and greensands but have also been found in limestones. Dimensions range to about 6 cm. in diameter; shape may be anything; and each has a more or less phosphatized nucleus around which are concentric laminae of collophanite and other substances. Nodules may become cemented to form a crust.

Phosphatic nodules are favorite places of attachment for benthos, and they may become extensively perforated by tubes of boring organisms. These organic structures tend to become phosphatized or filled with phosphate. The range in color is from brown to black, with a greenish shade on the surfaces of some.

In the present sea the phosphatic nodules are rarely pure and in many cases are merely calcium phosphate-cemented parts of the sediments. The source of a part of the phosphorus is generally assumed to have been that contained in shells and other organic matter. This phosphorus was placed in the colloidal state by action of organic decomposition products and finally precipitated around various nuclei in the sediments. Is not certain that the phosphorus was derived from organic matter. Some, perhaps all, may have been directly precipitated from sea water. Marine currents may work these sediments, remove the parts not phosphate, and thus concentrate phosphate. This may take place where a bottom has attained the base level of deposition.

In addition to the phosphatic nodules that form on the sea bottom, it is known that they have been formed in very shallow waters in coastal swamps and lagoons and perhaps in other swamps.

PRIMARY GUANO

Primary guano consists of excrements and remains of dead animals that accumulate on areas that are sufficiently dry to retard bacterial decay and are suited as dwelling places for vast numbers of animals, chiefly birds. These conditions are best realized on islands in the drying belts of the trade winds, where sea birds congregate in vast numbers and are free from predatory animals. The thickness of accumulation varies with the islands and also on the same island. Thicknesses as great as 6 meters have been formed.

Fresh guano is a mixture of many things, and its composition is in process of continuous chemical change from the beginning of accumulation. Change is hastened by high temperature and occasional rain, and the more volatile and soluble substances are evaporated or removed in solution almost as rapidly as produced. Chemical reactions affecting the original excremental and skeletal matter produce minerals that are characteristic of guano, among which are phosphates, nitrates, and oxalates. Many of these are removed shortly after formation, and the deposit ultimately passes into secondary or stone guano.

Some guano is formed in caves from the excrements and skeletons of bats. This is not greatly different from that formed on bird islands.

As guano deposits have commercial importance for fertilizer, those of any extent and thickness have largely been removed. None older than the Recent has been recognized in the geologic column.

Secondary Phosphates

LEACHED PHOSPHATES

A primary guano deposit continues in chemical change and removal, and in time through leaching and the destructive work of bacteria much is lost by solution and evaporation. Nitrates, oxalates, urates, ammonium phosphates, and other soluble constituents are thus removed, and only the less soluble constituents remain. Repeated collapses of the honeycombed guano and cementation of the fragments ultimately result in a breccia that is much richer in calcium phosphate than the primary guano. This is without stratification; much consists of agatelike particles; and the color is generally light. This is secondary, or stone guano. Comparison with analyses of primary guano shows losses in nitrates, urates, oxalates, ammonium phosphates and other very soluble constituents, and a large relative increase in calcium phosphate and other less soluble substances.

Any rock containing phosphates may have the same experience. Limestones containing small percentages of calcium phosphate may lose the carbonates and concentrate the phosphate, and under favorable conditions the residue may attain considerable thickness. Some calcium phosphate is also likely to be lost. This may be completely removed or deposited in underlying rocks. Deposits of residual phosphate are not uncommon and have been described from Kentucky, Tennessee, Belgium, France, and elsewhere.

A deposit of somewhat questionable classification may be made on bottoms overlain by waters so rich in carbon dioxide that previously deposited carbonates are removed and a residue composed of phosphate remains. No certainly proved occurrence of this kind is known, but thin layers of phosphate in the geologic column may be of this origin.

PHOSPHATIZED ROCKS

Phosphate deposits that are being leached may lose phosphate that is later deposited in underlying rocks. Rocks underlying primary guano are known to have become completely phosphatized to depths of more than a meter, and the same thing is also known to have taken place in the leaching of phosphatic limestones in Florida and elsewhere. In most instances the rocks replaced are limestones and coral-reef rock, but on Navassa Island in the West Indies the phosphatized rock was an andesitic lava, and the phosphates formed were those of iron and alumina instead of the common calcium phosphate.

DETRITAL PHOSPHATE DEPOSITS

Destruction of phosphate deposits by physical methods forms sands and gravels as do other rocks, and occasional deposits of phosphatic sands or gravels are thus made. Such are the "River Pebble Phosphates" of Florida, the "Land Pebble Phosphates" (river gravels on stream terraces) of the same state, basal phosphatic gravels in the Rocky Mountain bedded phosphorites, and phosphatized Jurassic fossils at the base of the Lower Cretaceous greensand of England.

The Phosphorus Cycle

Phosphorus released from igneous rocks ultimately may reach the sea and become widely diffused. It is utilized by organisms from the beginning of transportation and in part placed in solid form in shells, teeth, and bones, and in part placed in the tissues of both plants and animals. The plants and animals die, some by the so-called natural death; others are killed by animals for food. Scavengers eat all accessible dead bodies. The result is that the phosphorus continues in organic circulation, which may be endless. Some phosphorus escapes in excremental particles that fall in places protected against scavengers, and in some instances both animals and plants fall upon bottoms where deposition is so rapid that entombment is immediate or on bottoms so highly anaerobic that destruction is not possible because of absence of scavengers. Such are places of poor circulation, as bogs, deep lakes, deep holes on the ocean bottom, and bays and sounds of tideless epicontinental seas. Under conditions of rapid deposition phosphorus remains widely disseminated, but on anaerobic bottoms ammonia, hydrogen sulphide, and hydrocarbons develop and form solutions that may dissolve calcium phosphate. Reactions take place with calcium carbonate, and calcium phosphate is deposited. Sediments in these places are black from contained organic matter and iron sulphide.

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SULPHUR OF SEDIMENTARY ORIGIN

Introduction

Native sulphur of sedimentary origin is present in recent and some ancient sediments as isolated particles, lenses, beds, and irregular bodies of various forms. Associated substances are calcium sulphate, calcium and magnesium carbonates, and organic matter. Most sedimentary sulphur is associated with argillaceous sediments. Sulphur is precipitated through oxidation of sulphates and hydrogen sulphide in solution and from colloidal sulphur occasionally present in waters, and it may form as a decomposition product of metallic sulphides. Organisms in some cases are concerned in the deposition.

Environments of Deposition of Sulphur

Sulphur is formed and deposited under at least four sedimentary environments, namely, (*a*) the surface and in the zone of weathering through decomposition of metallic sulphides, (*b*) about springs, (*c*) in waters low in oxygen with reducing conditions and anaerobic bacteria, and (*d*) in bodies of water into which are brought volcanic gases containing sulphur. The deposits of group (*a*) are epigenetic; the others are syngenetic.

a. Decomposition of metallic sulphides may yield small quantities of free sulphur that takes the form of efflorescences or forms crystals. The free sulphur in some geodes may be of this origin. Decomposition of pyrite or marcasite in the Cheyenne sandstones of Kansas yields sulphur that appears in powdery form at the places of decomposition. Somewhat related and perhaps best associated with this group is the sulphur that is *present as thin films on gypsum in Permian and Upper Carboniferous strata of the Trans-Pecos Region of Texas.* There are indications that the gypsum is secondary and was formed through alteration of limestone. It has been suggested that the sulphur was formed from reduction of gypsum, but it may have been formed through oxidation of hydrogen sulphide which is stated to be present in circulating waters.

b. Native sulphur is a not uncommon precipitate about springs whose waters contain hydrogen sulphide, sulphates, and colloidal sulphur. The sulphates are thought to be reduced by bacteria to hydrogen sulphide, of which the oxidation produces water and colloidal sulphur, which may then be flocculated and precipitated. It may remain in suspension if the waters contain colloidal sulphur, sulphuric acid, and sodium sulphate in the proportions of S, 2.60 to 2.79 per cent; H_2SO_4, 6.43 to 7.00 per cent; and Na_2SO_4, 3.75 to 3.92 per cent. Any increase in the sulphur or decrease of the other constituents leads to precipitation. Deposits of sulphur of this origin should be small.

c. Bodies of water deficient in oxygen below the surface and containing dissolved sulphates may become filled with hydrogen sulphide produced by bacterial reduction of sulphates. Oxidation of the hydrogen sulphide, for which another group of bacteria may be responsible, leads to separation of sulphur. This method of formation of sulphur evidently has extensive occurrence in a small way, and Trask and Wu have shown that most marine muds studied by them contain free sulphur in quantities ranging from 22 to 104 parts per 100,000.

The sulphur deposits of south central Sicily have been postulated to have been formed in this way (Hunt). This sulphur is in Tertiary strata composed of gypsum, limestone, and bituminous clay. Some of the strata are typically marine. The deposits are more or less disconnected, and each is said to have basin form. Celestite, aragonite, and gypsum are associated with the sulphur. The sulphur-bearing zones range from 1 to 30 meters thick with the average 3 to 4 meters, consist of layers of pure sulphur in marly clay or limestone, gypsum, and some bituminous matter, and the zones are separated by bituminous shales. Sulphur layers are less than 2.5 cm. thick and constitute about half the thickness of the associated strata in each zone. Sulphur layers are known to extend for distances greater than 200 meters.

Some are cross-laminated, and local erosion during deposition was shown in one sulphur layer that was disconformably overlain by clay. Hunt suggests that deposition took place in more or less disconnected lagoons that did not have open connection with the sea, but occasional entrance of the sea is shown by marine fossils in some beds of the sequence. The gypsum evidently was deposited at times of poor connection with the sea, and the high salinity at those times eliminated the sulphate-reducing bacteria held responsible for the precipitation of the sulphur, *Spirillium desulphuricum* ceasing its activities when the concentration attains 3 per cent and *Microspira aestuarii* almost ceasing when the concentration exceeds 6 per cent. The sulphates were reduced to sulphides with hydrogen sulphide as the ultimate product. Rising upward, this was oxidized into water and sulphur, bacteria taking part in and perhaps being essential for the oxidation. These changes are shown in the following equations:

$$\begin{aligned} CaSO_4 + 2C \text{ (bacteria)} &= CaS + 2CO_2 \\ 2CaS + 2H_2O &= Ca(OH)_2 + Ca(SH)_2 \\ Ca(OH)_2 + Ca(SH)_2 + 2CO_2 &= 2CaCO_3 + 2H_2S \\ H_2S + O \text{ (bacteria)} &= H_2O + S \end{aligned}$$

The calcium hydrosulphide coming in contact with free sulphur might form a calcium polysulphide and liberate hydrogen sulphide as follows:

$$Ca(SH)_2 + 4S \rightleftarrows CaS_5 + H_2S$$

As this reaction is reversible, any hydrogen sulphide coming in contact with the polysulphide would precipitate sulphur, thus at times producing a larger precipitate than usual.

The hydrogen sulphide might also be oxidized through biologic agencies, perhaps, in the upper and oxidizing waters of the basins of deposition to sulphuric acid. This would react with any calcium carbonate in solution to form calcium sulphate, and this in the lower reducing waters might again become reduced to calcium carbonate and hydrogen sulphide. The range and environments of reactions are shown in Fig. 47.

In addition to the two bacteria named, *Proteus vulgaris*, *Bacterium mycoides*, *B. hydrosulfuricum*, and *Vibrio hydrosulfureus* function in processes like those outlined.

Not all agree that the origin of the Sicilian deposits was as described above, and Sagui has postulated a volcanic origin for the hydrogen sulphide and explains the gypsum as resulting from reaction between calcium carbonate, water, and sulphur.

The sulphur deposits of Sicily in accordance with the preceding (excluding Sagui's suggestion) are postulated to have formed something as follows: Lagoons at times were open to the sea, and normal marine sediments were deposited. At other times the connections were restricted, there was poor circulation with development of sulphur bacteria, and sulphur was deposited. Occasional restoration of circulation stopped sulphur deposition and permitted deposition of calcium carbonate. At times there was flooding of the places of sulphur deposition by black mud. Extreme restriction of the connections increased concentration of the waters of the lagoons and led to precipitation of calcium sulphate. Thus sequence of events with variations is postulated to have been repeated several times.

Bedded deposits of sulphur in Russia somewhat similar to those of Sicily have been described by Murzaiev. He favors an origin due to biochemical reduction of sulphates in lagoons that have a lower reducing zone and an upper oxidizing one, similar to the diagram of Fig. 47. Ahlfield, who described a deposit of bedded sulphur in Rimagna, Italy, reached a similar view for the origin of the deposit studied by him. He held that there was biological reduction of sulphates in a lower reducing zone and oxidation of the hydrogen sulphide produced in an upper oxidizing zone. The views of Murzaiev and Ahlfield are essentially the same as the view of Hunt.

d. Sulphur gases are escaping from crevices on the bottoms of some Japanese lakes. Sulphur is precipitated on contact with the water and deposited as hollow, variously shaped particles of 0.2 to 3 mm. in diameter. Deposits of this origin may attain considerable dimension.

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Sedimentary Deposits of Barium

Barium minerals of sedimentary origin do not make rock masses, and occurrences are limited to mineral particles, aggregates, and nodules. Most are barite ($BaSO_4$), but there is some witherite ($BaCO_3$).

The quantity of barium in surface waters is small, and often it is not detectable. It has been detected in the waters of the sea. Locally, in the waters of some springs, wells, and mine waters it is an important part of the solids in solution. Pipes carrying waters from some English coal mines have become filled with deposits of which barium minerals were the chief constituents, and oil wells in Kansas and Texas are known to have yielded waters that precipitated barite.

The barium in natural waters is acquired from the rocks through which the waters circulate, and it seems to be transported mostly in the form of chloride and bicarbonate. It may be precipitated as the sulphate by reaction with sulphate waters, and the carbonate may be deposited on loss of the carbon dioxide from bicarbonate. Barium bicarbonate in solution may be precipitated as the sulphate at places of contact with calcium sulphate or oxidizing iron sulphide. The occasional occurrence of barite in limestone and in the cement of sandstone may be due to this cause. Organisms also play a part in precipitation, as Schulze and Chevotiev have observed granules of barite in the bodies of some rhizopod protozoans.

Barium is present in detectable quantities in many marine deposits. There is a little in marine manganese nodules, and small more or less spherical nodules containing 75 per cent barite have been dredged from a depth of 675 fathoms off the coast of Ceylon. This barium seems to be of syngenetic origin. Barium is deposited as incrustations about springs, and this has been described from springs in Delta County, Colorado. A brine spring in a mine at Lautenthal in the Harz Mountains of Germany deposits barite from reactions of the barium chloride in the spring waters with the magnesium and zinc sulphates in the mine waters. Strontianite is also precipitated, but in much smaller quantity. Springs in British coal mines deposit barite and strontianite because of similar reactions.

Barite is a part of the cement of some sandstones, as in parts of the Chester series of Indiana, the Triassic of England, and the Nubian sandstone of Egypt. This barite is thought to be epigenetic.

Barite "concretions" are crystal aggregates that are usually found in sandstones and to some extent in sandy clay- and siltstones. These are disk-shaped, locally known as "barite dollars"; rosettes, termed "petrified roses"; or irregular. "Petrified roses" are tabular crystal aggregates of barite. They have been found in England, Russia, Egypt, Oklahoma, and elsewhere. They are epigenetic and in forming did not displace or replace the enclosing rock but included this in the aggregates as they formed. They usually are high in silica, which may exceed 50 per cent, and there are usually small quantities of alumina and ferric oxide. Barite disks have been described from England, Nebraska, and Egypt. Those in the Lias of England are irregularly distributed along bedding planes and have maximum diameter of 15 mm. and maximum thickness of 2 mm. Upper and lower surfaces have radial markings. Involved intergrowths and nodules are associated with the tabular forms in Egypt, and dimensions range to 70 mm. in diameter. "Barite dollars" are epigenetic, and they cemented into themselves the sands in which they formed. Thus they have a high content of silica.

Oölites and pisolites of barite have been obtained from oil wells in the Batson and Saratoga fields of Texas under conditions that showed they had formed in the wells. Some contained nuclei of pipe scale. Crystals of barite are not uncommon in geodes and cavities of shells. These are of epigenetic origin.

Reference

TWENHOFEL, W. H., "Treatise on Sedimentation," pp. 581–584, 1932.

SEDIMENTARY DEPOSITS OF STRONTIUM

Strontium minerals of sedimentary origin are not known as rock masses but as individual minerals and mineral aggregates of celestite ($SrSO_4$) and strontianite ($SrCO_3$). Neither is common. Strontium is acquired by natural waters from the rocks through which they circulate, and strontium chloride, strontium sulphate, and strontium carbonate may be taken into solution as they are encountered and transported in these forms, except that the carbonate needs to be placed in bicarbonate form. The element ordinarily is present in natural waters as mere traces, and only in rare springs and wells is the quantity in solution of any importance. Analysis of waters of a mine spring at Lautenthal in the Harz Mountains of Germany showed 0.899 grain of strontium chloride per liter and nearly three times as much barium chloride. Water in a gas well in Washington County, Pennsylvania, contained strontium to the amount of 1.31 per cent by weight of the solid matter in solution. It may be detected in ocean waters.

Precipitation of strontium sulphate, or celestite, takes place if waters containing strontium chloride or other soluble salts of strontium mingle with other waters containing sulphates or sulphuric acid. Strontium carbonate is precipitated from the bicarbonate through loss of carbon dioxide.

Some strontium is in organic circulation, as it is known to be present in the ash of seaweeds, and it has been detected in the shells of certain mollusks and corals; the radiolarian *Podactinelius* and the actipylean radiolarians, in general, have shells almost entirely composed of strontium sulphate. Samoilov has suggested that organic precipitation is responsible for the presence of extensive distribution of celestite in Turkestan.

Only traces of strontium are detectable in modern marine sediments, except in the manganese nodules where small quantities are present. Celestite and less commonly strontianite are deposited as incrustations about some springs, as at the Lautenthal mine spring, where celestite is associated with barite, and at some springs in Sicily, where the associated minerals are sulphur and gypsum.

Isolated crystals and crystal aggregates occur disseminated in various rocks and cavities. Crystals of celestite are present in dolomite in Monroe County, Michigan, and parts of the dolomite contain 14 per cent celestite. Celestite is associated with sulphur in the lower part of this dolomite. Celestite occurs in pockets and thin layers in limestone in Schoharie County, New York, and in Oneida County, New York, both celestite and strontianite are present in geodes in the Clinton limestone. Both minerals are in limestone of Jefferson County, New York, and there is abundant celestite in dolomite near Syracuse, New York. A cave in limestone lined with crystals of celestite was found in 1897 near Put In Bay on Lake Erie. Nodules and pockets of celestite are in the Glen Rose limestone near Austin, Texas, and there is also some strontianite. Celestite masses up to the weight of 100 lb. have been reported. Celestite fills cavities 0.9 meter wide and 7.5 to 17.5 cm. high at Cedar Cliff, Mineral County, West Virginia, and it serves as a cement in parts of the Oriskany sandstone of western New York. Beds and lenses of strontianite are interstratified with lacustrine limestones and shales near Barstow, California, and in the Avawatz Mountains in San Bernardino County, California, are beds of celestite interstratified with lacustrine clay, gypsum, and salt. The aggregate thickness of the celestite beds is 22.5 to 24 meters, and in addition to strontianite they contain gypsum, iron, and manganese nodules. Celestite and barite occur together in a bituminous limestone of Transylvania, and the Nummulitic limestone of Egypt contains crystals of celestite of which some contain fossils and others fill fossil shells.

Except for the incrustations about springs, which are syngenetic, most occurrences are epigenetic. It has been suggested that the celestite and strontianite in the lake deposits of California are syngenetic, but it is said that the facts are more in harmony with replacement of beds of carbonate. The facts seem to indicate that replacement was brought about by cold waters, and thus the deposits are of sedimentary origin.

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TWENHOFEL, W. H., "Treatise on Sedimentation," pp. 584–586, 1932.

FELDSPAR OF SEDIMENTARY ORIGIN

In recent years it has been shown that feldspar of sedimentary origin has wide distribution in the geologic column. So far as known, occurrences are limited to small crystals in other sediments, particularly limestone, but a few occurrences are known in sandstones, shales, and siltstones (Gruner and Thiel). Authigenic feldspar has now been found in unaltered formations of every geologic system from the Pre-Cambrian to the Tertiary. In the Watertown dolomite of the Pre-Cambrian of the International Boundary of the Rocky Mountain Region, Daly states there are places where authigenic feldspar composes 40 per cent of the rock. The feldspars so far found consist of albite and orthoclase, and perhaps microcline in the Triassic of northeast Ireland. Crystals are more or less euhedral and most measure between 0.01 and about 0.08 mm. in length and with thickness somewhat less. Crystals 1 to 2 mm. long have been found in the Cuddepah limestone of India (Spencer), and limestone on the Island of Rhodes has yielded crystals of feldspar with lengths of 2.5 to 3 mm. Crystals of microscopic orthoclase have been found in an ammonite shell; carbonaceous matter has been found in authigenic crystals of albite; and albite crystals containing radiolarians have been found in Eocene rocks of Pavia. It is possible that the last is hydrothermal.

Nothing is known as to the conditions of origin of sedimentary feldspar, but the conditions can hardly be very peculiar, as the feldspars are present in greater or less abundance in nearly every limestone that has been carefully studied. Cayeux has called attention to the absence of feldspar in the Cretaceous strata of the Paris Basin that contain glauconite. Spencer noted that the feldspars in the Cuddapah system are in calcitic but not in dolomitic limestones, but feldspar of sedimentary origin has been found elsewhere in dolomitic limestones. Lory noted the common presence of doubly terminated quartz, pyrite, and bituminous matter with sedimentary orthoclase and isomorphic quartz, and mica with sedimentary albite. According to Grandjean, lacustrine limestones do not seem to contain authigenic feldspar, and de Lapparent found authigenic feldspar in the Flysch of the Alps in limestones composed of granules, for the origin of which algae of the *Girvanella* type were considered responsible.

It is questionable if there are any genetic relations in the various associations given above. If feldspar does not form in lake sediments, the occurrence of feldspar becomes a valuable criterion for differentiation of sediments of fresh and salt waters.

The general opinion seems to be that the authigenic feldspars grew in calcareous muds; otherwise the euhedral character would be almost impossible of explanation; but Reynolds thought that the feldspars of northeast Ireland might have been formed after the consolidation of the enclosing rock. The temperature of formation could not have been otherwise than low, and Daly suggested that it might have been as low as 70°C. (158°F.). It seems very doubtful that any temperature so high ever obtained in some of the occurrences.

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SPENCER, E., Albite and other authigenic minerals in limestone from Bengal, *Mineralog. Mag.*, 20 (1925), 365–381.

SEDIMENTARY MICA

It has been known for a long time that minerals belonging to the mica group form in argillaceous sediments and soils. The minerals have been described as resembling muscovite and were designated hydromica, sericite, and a sericitelike mineral. German writers have applied the name of "glimmerton." The minerals are said to have different composition from any described mica, and to form authigenetically in argillaceous sediments soon after deposition and in soils as they form or alter by weathering. Dimensions are stated to be commonly less than 1 μ in diameter and to have a usual range in dimension between 0.1 and 0.6 μ. Composition is stated to approximate $2K_2O \cdot 3MO \cdot 8R_2O_3 \cdot 24SiO_2 \cdot 12H_2O$, where M equals ferrous iron or magnesium and R aluminum or ferric iron. It has lately been proposed to give to this mica constituent of the argillaceous sediments the name of *illite* (Grim, Bray, and Bradley), the name to be used in a general sense for the authigenic micas in argillaceous sediments rather than as a specific mineral term. Ross and Hendricks recently pointed out that Mallard had proposed the name of bravaisite for this substance in 1878.

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SELENIUM

Selenium, one of the rarer elements, is an important minor constituent in some sedimentary deposits. It has been studied rather extensively in South Dakota where it is present in the Niobrara and Pierre divisions of the Cretaceous system. Moxon, Olson, and Searight found the selenium content in these formations to range as high as 45 parts per million, to be common with 20 to 30 parts per million, and to be as high as 113 parts per million in a bentonite layer in the Niobrara. The selenium is taken up by plants and apparently placed in soluble form by what are known as converter plants, among which a species of the pea family, *Astragalus racemosus;* an aster, *A. multiflora;* a genus of the mustard family, *Stanleya;* and a few others are important. These concentrate the selenium in their tissues with a range to 4,800 parts per million and a common content of 1,000 to 3,000 parts per million. *Astragalus* seems to be the most important converter plant. Whatever the original source of the selenium and however it was carried to the ocean, Byers and associates have found that it is rapidly precipitated in the presence of colloidal ferric oxide as an insoluble iron compound and thus enters the sediments. It may be removed on the weathering of these or remain as a constituent of the soil from which it may be taken by converter plants. Death of the plants may return the selenium to the soil or yield it to ground water. If the plants are eaten by animals, they may produce "alkali disease" or selenium poison (Moxon) and kill the animals. This will return the selenium to the soil or ground water. Some selenium may also be released before death in the excretions, and some may also be released by the animals in volatile form. Obviously the sedimentary cycle of selenium is like that of the many other substances that are utilized by organisms.

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SEARIGHT, W. V., and A. L. MOXON, Selenium in glacial and associated deposits, *Agric. Exper. Sta., South Dakota College of Agric. and Mech. Arts. Tech. Bull.* 5 (1945), 1–33.

SEDIMENTS OF METEORIC ORIGIN

Deep-sea sediments contain particles referred to meteoric origin. A liter of red clay may contain 20 to 30 black magnetic spherules and a half dozen brown spherules of crystalline structure. Other sediments of more rapid deposition rarely contain these substances. The exteriors of the black spherules are composed of magnetite, and the centers are mostly iron or iron alloy. Surfaces are smooth, and diameters rarely exceed 0.2 mm. The brown spherules resemble the chondrite variety of meteorite. They contain silicon, have a range of color from yellow to brown, a pronounced metallic luster, and a striated surface. They are somewhat larger than the black spherules and average about 0.5 mm. in diameter. No meteoric particles are known from the geologic column unless the moissanite (Ohrenschall and Milton) identified in rock of the Mid-Continent is of this origin. It is generally considered that the moissanite was accidentally introduced into the sediments after acquirement.

Reference

OHRENSCHALL, R. D., and C. MILTON, The occurrence of moissanite in sediments, *J. Sed. Pet.*, 1 (1931), 96–99.

SEDIMENTARY ZEOLITES

Phillipsite, analcite, and apophyllite are three zeolites that have some distribution in sediments. Phillipsite is widely distributed in deep-sea sediments, and analcite and apophyllite are present in the Green River shales of the northern Rocky Mountains of the United States.

Phillipsite is a hydrous silicate of potassium and calcium, whose composition in some cases may be expressed as $(K_2 \cdot Ca)Al_2Si_4O_{12} \cdot 4\frac{1}{2}H_2O$. Crystals are microscopic and are formed as products of diagenesis in deep-water clays and oozes of the Indian and Pacific oceans. Murray and Renard ascribed formation to decomposition of volcanic materials. Some red clays are said to be composed of 20 to 30 per cent phillipsite, but the percentage is usually small.

Analcite has the formula $Na_2O \cdot Al_2O_3 \cdot 4SiO_2 \cdot 2H_2O$, and it forms more or less persistent thin beds in the Green River shales (Bradley). The mineral is in the form of euhedral crystals that range in dimension to nearly 2 mm. All crystals are clouded with dustlike inclusions, and they are cemented in a matrix of chalcedony. The analcite crystals vary in proportion to the matrix in different beds, and where the analcite crystals are few the rock has a tuffaceous appearance. The beds contain occasional crystals of apophyllite.

Isolated crystals of analcite and apophyllite are also present in many of the associated oil-shale beds, and some consist of as much as 16 per cent by weight of analcite and more than 1 per cent of apophyllite.

Bradley explains the zeolites in the Green River shales as results of reactions between salts in solution in the waters and dissolution products of volcanic ash. Various volcanic minerals and a little volcanic glass are in association with the zeolites and support the interpretation. The deposits are probably penecontemporaneous.

Merritt and Ham have lately described natrolite in a sedimentary deposit in the Wichita Mountains of Oklahoma. They suggest that the mineral is a product of diagenesis. Sedimentary heulandite has been found by Gilbert and McAndrews in Upper Miocene in Santa Cruz County, California. It is a minor constituent and forms less than 1 per cent of the rock. Occurrence is in the form of idiomorphic crystals and crusts on other minerals. The authors consider it certain the heulandite crystallized after deposition of the sands and at low temperature.

Other zeolites resulting from decomposition under conditions of low temperature are chabazite, stilbite, and laumontite. These are of epigenetic origin, but there seem to be no reasons why they might not form in sediments as penecontemporaneous minerals.

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Merritt, C. A., and W. E. Ham, PreCambrian zeolite-opal sediments in the Wichita Mountains, Oklahoma. *Bull. Am. Assoc. Pet. Geol.*, 25 (1941), 287–299.

SEPIOLITE (MEERSCHAUM)

Several thin beds of sepiolite or meerschaum ($2H_2O \cdot 2MgO \cdot 3SiO_2$) are present in the Green River formation in Duchesne County, Utah. The associated strata consist of chocolate-brown shale containing an abundance of calcite casts after glauberite. No bed exceeds 1 cm. in thickness. The sepiolite is explained as due to deposition from water of magnesia and silica with the silica perhaps in the form of gel. The mineral contains small quantities of aluminum, ferric, calcium, and titanium oxides. The deposits are syngenetic. The mineral is also present in other parts of the world.

SEDIMENTARY TOURMALINE

Tourmaline is known as a sedimentary product because of enlargement of rounded particles of the mineral. It has been described in North America from Lower Devonian sandstones of Michigan (Alty), the Oriskany sandstones of the Appalachian Region (Stow), and the Whirlpool sandstones of the Niagara Gorge (Alling).

The tourmaline from the Lower Devonian of Michigan consists of growths on rounded particles, the growth always taking place at the end of the *c* axis. The added tourmaline is in optical continuity with that of the original particles, but has a much lower index of refraction. The tourmaline in the Oriskany sandstone is of the same character and consists of a jagged growth on well-rounded particles at one end of the *c* axis.

Tourmaline particles with authigenic growths have rather general distribution in the Oriskany sandstones from Virginia to New York, so that it is not possible to ascribe them to any local conditions.

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ZINC, LEAD, AND COPPER COMPOUNDS

Mine waters carry compounds of zinc, lead, copper, and some other elements and locally make small deposits. Deposits of lead, zinc, copper, and some other elements undergo enrichment in the zone of weathering through cold-water transportation and deposition of oxides, carbonates, sulphates, and, rarely, other compounds of the metals. Mine waters of the Joplin District of Missouri deposit zinc sulphide, and many geologists consider that the deposits of lead and zinc sulphides in the Tri-State District and southwestern Wisconsin were deposited by cold waters that acquired lead and zinc widely distributed in the surrounding rocks. Versteeg reports crystals of galena and sphalerite present in shale, and replacements and filling fossils in the Cuyahoga limestone of Ohio and sphalerite is also present in the Pottsville formation.

Pure copper and many of its compounds have been deposited by sedimentary processes. Copper sulphides are oxidized, carbonated, and dissolved in upper parts of deposits and carried downward and deposited as cuprite (Cu_2O), tenorite (CuO), malachite ($CuCO_3 \cdot Cu(OH)_2$), and azurite ($2CuCO_3 \cdot Cu(OH)_2$). Chalcocite (Cu_2S) has been made in the laboratory in cold solution, so that it is possible for it to be a sedimentary product. Bronze articles about Swiss lakes have become coated with chalcopyrite ($CuFeS_2$) showing that this mineral may also be formed by sedimentary processes.

The Kupferschiefer (Trask) (Permian) of Germany contains bornite, chalcocite, and chalcopyrite that have been referred to a sedimentary origin. These minerals are associated with, and enclosed in, dark shales. Zinc and lead minerals and native silver are also present. The minerals are generally considered syngenetic with the enclosing sediments, and precipitation of the copper compounds has been referred by some to organisms or to chemical reactions connected with decay of organic matter.

Native copper has been found in unaltered sediments as films near Enid, Oklahoma, and as impregnations of fossil wood in the Permian of Texas. A peat bog in Wales contains 3 per cent copper. Thin beds of copper in black muck of a bog near Cooke, Montana, have been described by Lovering. The copper is in the form of spongy masses that range from mere specks to 2.5 cm. in diameter. The copper is thought to have been brought to the bog as cupric salts in solution and to have been precipitated by reduction of copper sulphate by microorganisms living in the materials of the bog. Experiments showed that these microorganisms were able to live in a 1/2,500 solution of copper and to precipitate metallic copper from copper sulphate solutions ranging from 1/10,000 to 1/50,000.

The circulation fluids of many invertebrates contain copper in the form of a copper protein known as hemocyanin. As most organisms are eaten after death, most of this copper remains in organic circulation; but if dead organic matter is buried, copper is introduced into sediments in this way. Burial may take place where bottoms are of such character that organisms cannot endure the conditions. The quantity of copper introduced in this way may be large, but as studies are limited, it is not possible to make estimates. Studies by Galtsoff and Whipple have shown that certain oysters are green because of the presence of a copper compound in a highly disassociated unknown state, and that this copper is in addition to that in the circulation fluids. The copper content of normal oysters ranges from 0.16 to 0.248 mg. per oyster, but in the green oysters the range is from 1.24 to 5.12 mg. or an average of over 3 mg., about twenty times as much. Considering the countless billions of organisms now in the sea and the many more that have lived during geologic time, it would be strange, indeed, if organisms had not been responsible for the deposition of considerable copper.

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TITANIUM MINERALS

Rutile, anatase, and the supposed mineral leucoxene, lately shown to be a microcrystalline form of rutile and anatase (Tyler and Marsden), may form in sediments under cold-water conditions. These minerals form from ilmenite under conditions of low pressure and temperature, thus accounting for the presence of perfect crystals in unaltered sediments.

Reference

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ZIRCON

Butterfield has described zircon with toothlike projections in the form of small pyramids attached to the prism faces. These projections seem to be of authigenic origin. More data are desirable.

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SEDIMENTARY GOLD

It is possible that, under certain conditions, waters of surface temperatures may transport and deposit nonclastic gold. It has been reported that gold in gravels and sands in Brazil (Friese) had been increased between a first and second working and that the gold added was different from the original. The view is advanced that gold is slowly attacked and removed by humic acids under conditions excluding oxygen. Experiments seem to support the view. The field evidence is said to indicate that deposition of the gold is related to plant growth. The diamond-bearing gravels of Minas Geraes are stated to be partly cemented by gold deposited from solution.

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CHAPTER XIV

STRUCTURAL FEATURES OF SEDIMENTARY ORIGIN

INTRODUCTION

Structural features of sedimentary origin include stratification, cross-lamination, unconformity, ripple mark, wave mark, soft-rock deformation, concretion, cone-in-cone, and others of less importance. Some of these features are important for the light they throw upon conditions of deposition, and knowledge of original positions may permit determination of reversals of strata and thus give the key to solution of complex structure.

STRATIFICATION

Stratification (Andree, Barrell), or arrangement in layers or beds, is the most distinctive feature of sedimentary rocks. It arises from variations in color, texture, dimensions of particles, and composition, or to temporary cessation of deposition that permits already deposited sediments to undergo some changes before renewal of deposition. A *stratum* is a unit of sediments that separates more or less readily from overlying and underlying units. All strata are lenses of greater or less areal extent, and each thins laterally to disappearance, but there may be passage into some other variety of sediment before a stratum disappears. Stratification planes separate strata, or parts of strata, in case a stratum is internally stratified. If stratification planes are 1 cm. or less apart, the sediments are said to be laminated and each thin unit is a *lamina*. Larger units in a stratum are termed *stratum layers*. Lamination may be parallel to bedding planes or at some angle thereto. In the latter case the term of cross-lamination is applied.

Stratification has been classified as *direct*, or *primary*, and *indirect*, or *secondary*. The former is that made when the sediments are first deposited. Indirect stratification develops if sediments already deposited are thrown into suspension and redeposited.

Stratification varies from very regular and even to the other extreme of highly irregular. The former is likely in sediments deposited in deep or very shallow quiet waters; the latter in sediments deposited in agitated waters. Many strata show no internal stratification. This may be due to continuous deposition of a single variety of sediment, but, more likely, to reworking of sediments by burrowing and scavenger organisms.

INCLINATIONS OF STRATIFICATION AND DIMENSIONS OF SEDIMENTARY UNITS

Each stratification plane tends to be parallel to the surface upon which deposition takes place, and each stratification plane in turn becomes a surface of deposition. In a long sequence of strata there is likely to be considerable divergence from parallelism of the stratification planes at the bottom and the top. Initial surfaces of deposition are not necessarily horizontal, and existing surfaces upon which deposition is taking place are undulatory and contain basins and elevations of divergent dimensions. Neither are surfaces made by deposition necessarily parallel to surfaces of deposition immediately below. The lower part of a surface of deposition is likely to receive thicker deposits than higher places, and each successive stratification plane may have less inclination than the preceding. There are many exceptions, and deposition may make a new surface more irregular than the preceding and also much steeper locally. Interruptions of deposition by erosion may create surfaces of great irregularity.

Some study has been given to maximum steepness of initial inclination. Thoulet stated that the steepest inclination on which it is possible for sediments to be deposited is 41 deg. Recent studies have shown that coarse-grained sediments and sediments with angular particles may be deposited under water at steeper inclinations than finer grained sediments or sediments composed of rounded particles. The highest inclination attained under water was found to be 43 deg. with mixed coarse-grained angular to subangular sands. The steepest inclination found for very fine grained sands was 33 deg. The maximum slope of deposition for medium-grained sands was found to be 38 deg. for angular and 35 deg. for rounded particles. Clay will not accumulate on slopes exceeding 30 deg. unless it is adherent to solid materials, and with clays there is always a tendency for thinning on the upper part of any inclined surface with corresponding thickening on the lower part of the same surface.

Many sites of deposition are steeply inclined. Bottoms seaward of coral and other reefs are often of steep inclination, many submarine fault and volcanic slopes exceed 40 deg., and slopes of 20 deg. are not uncommon. Deposition of sediments often creates steeply inclined surfaces, particularly near the mouths of rivers flowing into quiet bodies of standing water. Slopes of this origin with inclinations as great as 35 deg. are known. Depositional slopes at the mouths of rivers flowing into the sea tend to be low and average less than 1 deg. for the Mississippi and ½ degree for the Rhine. This is due to the spreading action of waves and currents. Inclination of planes of stratification in stream deposits may be high and of great irregularity because of the prevalent cut and fill in streams.

Many sediments have high initial inclinations, and strata initially inclined at angles of 5 to 10 deg. are common. Ancient reef formations have inclinations exceeding 40 deg. The highest initial inclinations are shown in travertine deposits, where initial inclination of stratification planes may be greater than 90 deg. These are indurated from the beginning, and it is thought that most sediments having initial inclinations exceeding 40 deg. may be considered to have been deposited in solid form coherent to the surface upon which deposition took place.

Units of stratification range from paper-thin to many meters. The general range is 5 to 50 cm. The thickness is not a measure of the rate of deposition or of time. A thin layer of sediments of any kind may have taken as long a time to deposit as many meters of the same or some other kind of sediment at a different time or place.

Dimensions of sedimentary units vary with the environments of deposition and the supply and character of composing sediments. Coarse-grained sediments, large loads, and conditions producing rapid decrease in competency make for great variation in thickness and limited areal distribution. Fine-grained sediments tend to have fairly wide areal distributions and somewhat uniform thicknesses.

Deposits of alluvial fans and cones and floodplains, channels, and deltas of rivers are likely to show much variation in thickness and areal extent of sedimentary units over very short distances, and the irregularities increase as environments change from deltas to floodplains to channels and to alluvial fans. This variation is excellently shown in such fresh-water formations as the Fort Union of Montana and Wyoming, which is an assemblage of lenses of clay, silt, sand, gravel, and coal. Some of the units have areal extent of only a few square meters and others extend over many square miles.

Sedimentary units deposited in the littoral environment and on the adjacent shallow-water bottom have great dimensional variation. A unit may extend for many miles along a coast and on other parts of the same coast another unit may extend for only a few meters. Similar dimensional variation may be expected about reef structures, except that all sediments are likely to be calcareous.

Dimensions of sedimentary units over the deeper parts of neritic bottoms are related to distance from land and depth of water, but areal distribution is greater than under other environmental conditions except those on bathyal and abyssal bottoms. There seems to be little variation in thickness, and the units are lenses of vast areal extent. The areal extents of units on bathyal and abyssal bottoms are extremely large.

Most geologists up to the end of last century largely studied shallow-water formations of epicontinental seas, and there grew up the concept of large areal extent of depositional units. And some are extremely large. The Prairie du Chien formation of the upper Mississippi Valley has areal distribution of many thousands of square miles, a thin bed of limestone in the Anticosti section known as the "Track Bed" has been followed for 75 miles with little change in thickness or character, and there are units in the Mid-Continent Pennsylvanian that extend with little variation for more than 100 miles.

But there are as many, if not more, marine formations that are known to have great variations in thickness. On the island of Trinidad a formation known as the Williamsville Clay changes in thickness from 12 to 60 meters in 5 miles, a marl formation from 15 to 450 meters in 2.5 miles, and a green clay from 15 to 195 meters. The Tertiary of the Pacific Coast is full of units of great variations in thickness.

It is believed that the generalization may be made that deposits of existing continental and shallow-water marine environments are of variable characters and thickness and that most units are lenses of various dimensions. It is believed that such has always been the case.

ORIGIN OF STRATIFICATION

The origin of stratification is clear in some cases, but in others it is not understood (Andrée). It is strange that so little is known of this most distinctive feature of sediments. Tractional transportation inevitably develops stratification because of variations in competency that are brought about by deposition, and suspended sediments become sorted and to some degree stratified because of differences in rates of settling. Probable causes of stratification are changes of weather, changes in climate, changes in competency of currents not closely related to weather, rise of sea level, growth of organisms, and settling of suspended sediments. Stratification at any place may be due to a single one or several of these causes. It has been suggested that entrapped gases may produce stratification in silts and clays (Birch).

Weather and Seasonal Changes

Changes of weather are considered responsible for some stratification. Rainy weather produces floods, which bring much sediment to sites of deposition, and thick and widespread deposits of clastics may be made. Times between floods produce thin deposits of particles of smaller dimension and perhaps different composition. Stratification due to changes of weather may be obvious in sediments deposited in land environments, but some may also be present in marine environments, as strong winds produce waves and currents that may transport sediments from shores and shallow waters to deep waters, thus bringing sediments that ordinarily are not deposited under those conditions. Much change of weather is related to seasons, and in any region certain seasons are apt to be rainy and stormy. Other seasons may have little rain and wind and thus little transportation of terrigenous sediments to sites of deposition.

Much stratification in limestone formations is due to thin clay partings between beds. This may be seen in the Silurian and Ordovician limestones of the Upper Mississippi Valley in which the clay partings are of almost microscopic thickness, in Ordovician limestones over the Cincinnati Arch, in Mississippian limestones over the Mississippi Valley, and other limestones in many parts of North America. The thin clay partings seem readily explainable as due to storms stirring bottom deposits and making waters turbid, or to heavy rains on the land causing streams to flood and carry muddy waters long distances from the shore. The clays would settle last and form thin layers to become the partings separating the calcareous beds.

Seasonal deposition of sediments is best shown in lakes receiving melt waters from glaciers. Rapid melting in summer releases relatively large volumes of clastic sediments to melt waters, and at any place in such lakes deposits of a certain degree of coarseness and thickness are made. As weather becomes colder the volume of melt water progressively decreases, and progressively finer sediments are deposited. Finally ice covers the waters, and colloidal materials held in suspension are gradually deposited, with the very finest last. Thus the coarser deposits of summer grade into the finer deposits of winter, and, as the latter represent those held in suspension for a long time, and as some oxidation may have taken place, the winter deposits tend to have darker colors than those of summer. Melting in the spring abruptly renews summer deposition, and thus the change from the finer sediments of winter to the coarser of summer is sharp. The composition of winter and summer layers is much the same, except that the former contain more clay and ferric oxide. A summer and an overlying winter layer form a *varve*, or the deposit of a year (Fraser, Kindle, Sayle). The thickness varies from a few millimeters to 30 cm. or more. A varve may be laminated, owing to variations in character and supply of sediments, variations in competency of melt waters, differences in melting from day to night, and differences in temperature of melt and lake waters. So far as is known, melt waters of glaciers must flow into freshwater lakes to produce varves. Experiments have indicated that the maximum salinity permitting formation of varves of this origin with coarse clay is about one-fiftieth that of normal sea water (Fraser). Higher salinities flocculate the clays, and nothing remains for the slow settling of winter. Temperature has influence, as the rate of settling of particles less than 0.5 mm. in diameter decreases with fall of temperature to 4°C., and the slowness of settling increases inversely as the diameters decrease.

Organic remains mixed with muds form black deposits known as *sapropel*, *gyttja*, or *dy*. These may be deposited during one season of the year and other sediments may attain deposition during another part, thus forming a varve. Such seems to have been the case in McKay Lake (Whittaker), near Ottawa, Canada, of which the sediments consist of alternate laminae of light-colored lime carbonate and dark-red organic matter without a trace of lime carbonate. Two laminae are about 0.5 mm. thick. Varves in the Green River (Bradley, 1929) formation range in thickness from 0.014 to 9.8 mm., the former in the richest oil shales, the latter in the fine-grained sandstones. Bradley (1931) and others have noted the probable occurrence of varves in various other deposits, of which some are marine.

Climatic Changes

It is well known that climatic changes have taken place, and appeal has been made to these to account for some stratification. Climatic changes, however, are gradual and extend over long periods of time, and hence little detailed stratification is probable. Differences in large units are possible. Gilbert appealed to the astronomical cycle of the precession of the equinoxes to produce climatic changes leading to deposition of several units, not single strata, in the Cretaceous of Colorado. Other explanations of the features observed by Gilbert are more simple.

Variations in Competency of Currents

Much variation in competency is due to weather. This has been considered. But there are other variations, and these exist even with constant discharge on a laboratory stream table, making it almost impossible to prevent lamination. In a larger way, competencies vary due to individual or combined effects of winds and tides without change of weather. These produce some stratification, as is very obvious on any seashore affected by strong tides and tidal currents.

Stratification Connected with Relative Rise of Sea Level

It is thought that considerable stratification is due to rise of sea level. Bottoms become built to the base level of deposition with cessation of deposition until conditions make possible a higher base level. This may be done most readily by relative rise of sea level. Each rise permits deposition of a thickness approximately equal to the rise of sea level, and a new stratum or group of strata is formed that is separated from the units below by the surface of the former base level of deposition on which the later deposits were laid down. This surface represents a break in deposition that may have been very short or very long. It is either a diastem or a disconformity. Much of the stratification in limestone sections like those of the Ordovician and Silurian of the Michigan Basin, the Ordovician of the Cincinnati Arch, and elsewhere may be due to conditions connected with rise of sea level.

Reef formations show crude or rude stratification, and some of this should be connected with rise of sea level. Upward building by coralline organisms ceases when building has been brought to sea level. Rise of sea level permits upward building to be renewed. Positions of stability of sea level should be indicated by places of lateral expansion of the reef structures and by reef debris.

Stratification Due to Organisms

Congenial bottoms are carpeted with organisms and organic debris. Changes of temperature, turbidity, depth, character of bottom, circulation or salinity may make a bottom uninhabitable for one assemblage of organisms and permit entrance of another assemblage. The result is stratification. That something like this has taken place in the past seems shown by the occurrence of a stratum filled with a certain organic assemblage and an overlying stratum with a different assemblage.

Settling of Suspended Sediments

Suspended sediments settle in accordance with dimensions, specific gravities and shapes. The largest, heaviest and roundest are deposited first and the smallest, lightest and least round particles last. If settling acts alone, there is produced a stratum that is coarsest at the base and finest-grained at the top. Color differences are likely accompaniments of the other differences. Colloidal particles in suspension in distilled water arrange themselves in stratified form while in suspension. Settling in this arrangement may produce stratification. It is not considered, however, that this is likely to take place in natural waters and that any stratification in the geologic column is due to this cause.

If salts are introduced in colloidal suspensions, lamination results. In the Liesegang experiment a glass plate was coated with gelatin impregnated with potassium bichromate, and a drop of silver nitrate was placed upon it. Concentric rings formed around the drop. These were closely spaced near the center and progressively wider toward the outside. Certain conditions made equally spaced rings possible. When a solution of ammonium carbonate is carefully introduced into an aqueous suspension of finely divided clay containing a moderately strong content of sodium silicate there is downward diffusion of the ammonium carbonate leading to flocculation of the silicic acid and production of alternate laminae of fine clay and clear gelatinous silica. Substitution of powdered red shale or powdered crystalline quartz for the clay produces similar laminae (Davis). It is possible that some lamination of fine muds and siliceous sediments was formed in this way.

Cyclic or Rhythmic Deposition

Essentially all deposition is cyclic or rhythmic (Moore, Stamp, Weller). A site of deposition has a certain depth. It receives deposits determined by the conditions, the water becomes shallower, different deposits are made, and ultimately the bottom is built to the base level of deposition. Rise of sea level initiates a new cycle, and, if other conditions have not changed, the sequence of events and sediments is repeated. A land is elevated, and sediments determined by the conditions are transported to the sites of deposition, the land is brought to base level, and the sediments record the reduction in elevation. Reelevation starts the cycle anew. There are also cycles due to weather, climate, elevation of the sea bottom, formation of deltas, etc., and in some way all sediments record cycles of different origins.

There are few geologic sections that do not show cyclic arrangement of strata to some degree, and in some sections cycles within cycles may be seen. Some cycles consist of alternations of coarse and fine sediments, others of limestones alternating with shales, and in a higher order there is cyclic arrangement of zones of limestones and shales with zones of shales, sandstones, or limestones.

Criteria have not yet been devised to determine conditions responsible for each form of cycle, and the tendency has been for each student to relate all cyclic arrangements to one or two causes. This is a serious error. Conditions of importance in producing cyclic arrangement of strata are the following.

Weather and Seasonal Changes

Floods produce deposits of coarse-grained materials over some sites where previously fine-grained sediments had been deposited; subsiding waters lead to return of deposition of fine-grained sediments. Every storm stirring bottoms of standing waters renders the waters turbid; the largest particles settle first, the smallest last. The next storm repeats the cycle. A storm sweeps sediments of some dimensions into deeper waters where they are deposited over sediments of smaller dimensions to form a cycle.

The most outstanding example of the seasonal rhythm is the varve already considered, and perhaps in the same class is the rhythmic arrangement of salts in the evaporite deposits of the Permian of Texas (Udden) and Alsace (Gale). The former consists of cyclic arrangement of laminae of anhydrite, and 1,737 laminae were counted in 396 mm. The latter has a layer of sodium chloride alternating with a layer of sylvite, the former assumed to have been deposited in summer, the latter in winter. Laminated sediments in the Tertiary of Burma have been assigned by Stamp (1932) to seasonal variations in the capacity of the rivers flowing into the basin of deposition. Each postulated annual layer consists of a coarser grained part deposited during a high-water period and a finer grained part deposited during a low-water period. Rubey has suggested that laminae of a twofold character in the Cretaceous of the Black Hills Region may represent seasonal rhythm. The Tertiary Green River (Bradley) shales contain rhythmic alternations of thin laminae of carbonate and organic materials that were postulated by Bradley to represent seasonal deposition, and Bradley and others have noted similar alternations of sediments in various parts of the column.

Climatic Cycles

The repeated continental glaciations of geologic time suggest climatic cycles of long duration, and the several advances and retreats of the last glaciation suggest others of less duration. The withdrawals of water from the sea to make the glaciers lowered sea level, and melting away of the glaciers raised sea level, thus producing cycles connected with movement of sea level. Other cycles are the 11-year interval of sunspot maxima, the Bruckner 35-year period, the 21,000-year period of the precession of the equinoxes, and the 91,000-year period of minimum and maximum eccentricity of the earth's orbit. These cycles probably have some effect upon sediments, but what they are is not known with certainty. Gilbert suggested that the precessional cycle may have been responsible for a rhythmic effect present in some Cretaceous strata of Colorado, and Barrell postulated that a cyclic arrangement of light-colored slates, containing considerable sand, and thicker bands of dark strata represent climatic cycles of quiet periods following others of storm.

Rhythmic laminations in the Monterey formation of California have been described by Bramlette and termed *rhythmites*, a name previously proposed by Sander. Bramlette reached no decision as to origin but thought that they probably were in some way related to climatic control.

Rhythms Caused by Movement of Sea Level

Sea level probably changes for many reasons, among which are withdrawal of water to make the ice of glaciers and melting as the glaciers disappeared, movement of the crust, deposition of sediments on the ocean bottoms, sinking of bottoms under load, and submarine volcanic activity.

Any rise of sea level leads to submergence of some land, and the submerged areas receive sediments of some character and some degree of coarseness. These pass upward into other sediments of some differences of texture and character until the bottom is built to the base level of deposition, which may be above sea level under some conditions and below it under others. The last unit in the sequence is either continental or marine. A rise of sea level starts the cycle anew. Cycles, termed cyclothems, described in Illinois and elsewhere, consist of continental and marine sediments and in complete descending development consist as follows (Weller):

Marine:

8. Shale with thin layers of limestone in the lower part.
7. Limestone.

6. Calcareous shale.
5. Black shale.

Continental:

4. Coal.
3. Underclay.
2. Sandy and micaceous shale.
1. Sandstone.

Unconformity.

The deposits seem to belong to the delta environment, and the unconformity at the base of the cyclothem was interpreted as indicating subaerial erosion followed by subsidence. Unless proved by the evidence, it does not seem necessary to postulate uplift to explain the unconformity, as the sites of deposition may have reached the base level of deposition and thus a surface of erosion would have been developed until covered by advancing continental deposits of a delta represented in the sandstone.

A cyclothem (Moore) from the Douglas and Shawnee groups of the Pennsylvanian of Kansas in ascending order consists of 1.5 to 3 meters of yellowish-brown limestone, 1.5 to 4.5 meters or more or light-colored shale, 0.6 meter of hard blue limestone in a single bed, 0.9 to 2.4 meters of dark-blue shale with black shale at the base, and 3 to 6 meters of light-gray unevenly bedded limestone in thin beds. This cycle seems best interpreted as due to intermittent submergence with some aiding conditions on the land. The Illinois cycle seems best interpreted as caused by intermittent subsidence with perhaps in some cases elevations of the sites of deposition. Wanless and Shepard seem to wish to explain cyclothems in Illinois as caused by periodic development of continental glaciation that lowered sea level as water was extracted and raised sea level as the ice melted. Although this may be the explanation of some cyclothems, there is no evidence in the sediments of Illinois that supports the explanation.

Cycles of sedimentation described by Stamp from the Tertiary of the London, Paris, and Hampshire basins seem to have been made by periodically rising sea level and begin with a basal conglomerate that is progressively younger landward and is succeeded seaward and vertically by marine sand and clay, and these in turn by sediments of a continental character until the deposits were built to the base level of deposition or a retreat of the sea brought it to that level. A new advance of the sea started a new cycle. Each new advance removed some of the deposits of the previous cycle producing the variety of disconformity that Stamp terms a *ravinement*. Each cycle is limited above and below by ravinements (Fig. 19).

Large cycles, termed circles or cycles of sedimentation which comprised an entire geologic system and extended over a large area, were once assumed. These cycles were assumed to have a sand base, followed in order by shale, limestone, shale and, if the cycle was completely preserved, by a sandstone top. This cycle would not be shown in deep-water deposits a considerable distance from land. It seems rather unlikely that cycles of this character can be established.

Cyclic arrangement of marine deposits may be made by intermittent rises of sea level as noted in the introductory paragraph of the topic.

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CROSS-LAMINATION

Cross-lamination, also designated *cross-bedding* and *false bedding*, is a common feature developed in the deposition of sands. It consists of arrangement of laminations transverse to the planes of stratification of the strata concerned. Cross-laminae of a bed generally end abruptly at the top, but, in general, they tend to become more or less parallel to the bedding planes below. In the initial stage of formation the tops of laminae also tend to parallelism to the surface of the cross-laminated unit, but in the continued development of new laminae this top is removed. Very coarse sands are likely to end abruptly both above and below. The bottom tangential part of a lamina is designated *bottomset;* the inclined part, *foreset;* and, if present, the top tangential part, *topset.* Angles of inclination have a wide range from the known maximum of 43 deg. to very small. Most are less than 30 deg. Inclinations are related to the quantity and coarseness of the sands and rates of deposition. Inclinations are higher with rapid deposition and large quantities of coarse sands. After deposition there may be some reduction of inclination because of compacting. Foreset lengths are largely inversely controlled by quantity of sands and directly by velocity of currents, large quantities producing short foresets, high velocities long foresets. Foresets may range in length to 30 meters or more; but about 3 meters is an average high maximum. A form of cross-lamination in the Caspar and Fountain (Knight) formations of Wyoming has been designated *festoon.* Deposition is said to have taken place in southwestwardly plunging troughs open at the lower ends in the form of thin laminae conforming to shapes of troughs. As these troughs were filled, others are said to have been formed by erosion. Troughs range to over 30 meters deep, 300 meters long, and 100 meters or more in length. Maximum slopes to the sides of the troughs are 10 to 15 deg. and rarely 25 deg. Many laminae are intricately crumpled owing to slumping following deposition. Knight considered the formations of marine origin, but this interpretation seems impossible. Some features suggest aeolian deposition and others deposition under fluvial conditions. Perhaps both agencies participated.

Cross-lamination is a valuable aid in solution of structure. Each lamina forms a curve that is concave upward. Topsets are rarely present, and thus each lamina is truncated at the top of a bed. Except in coarse sands, each bottomset is tangent to the bottom. It is thus extremely easy to determine whether or not strata have been overturned.

Cross-lamination develops in all environments in which sand is deposited, and whether or not it is present depends upon the velocities of currents and supplies of sand.

Cross-lamination in streams always has a component of inclination downstream except in streams affected by tides, in which there may be cross-lamination upstream and perhaps a little local cross-lamination upstream may arise from eddies. However, streams meander, and this may lead to the development of considerable cross-lamination up-valley. Cross-lamination on alluvial fans, cones, and deltas has inclinations that fan out from the stream channels. Cross-lamination in lakes and the sea should be inclined in several directions in the strata of a sequence, as currents may move in one direction in the morning and in another in the afternoon. Cross-lamination in dunes follows the general pattern of *barkhans*, that is, deposition in troughs and on ridges of various dimensions, but plunging at any place in the same general direction. The laminations parallel the lee sides of the dunes (Fig. 61).

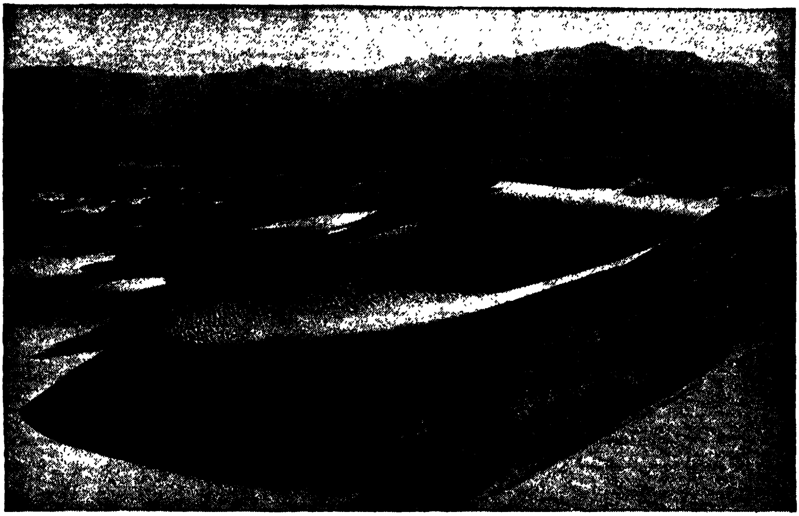

FIG. 61. Dunes south of Palm Springs, California. The inclinations of the laminations in the dunes are parallel to the surfaces of deposition and the directions of inclination may range through more than 90 and probably less than 180 deg. The anastomosing form of ripple mark characteristic of wind current ripples is beautifully shown. Both the dune and the ripples show that the currents move from right to left. It should be noticed that the directions of inclinations of the laminations in the dune are not necessarily in the same directions as the wind that transported the sands. [*From the Palm Springs Pictorial* 9 (1938–1939).]

Cross-lamination is made by water and wind. Both have the same general features previously given, in that the laminae in both are tangential to the surfaces of deposition at the bottom and are truncated at the top. Both have laminae concave upward. Inclinations of foresets do not seem to be different and each may have long foresets. There are differences in the form of the cross-laminated units. The bounding planes of aqueous cross-laminated units under most conditions are approximately parallel, whereas in aeolian cross-lamination the truncating planes are more likely than not to have considerable inclination to planes of deposition, so that units of aeolian cross-lamination are wedge-shaped when observed in two dimensions

(Fig. 62). Aqueous cross-lamination with short foresets occasionally has wedge-shaped units over small areas.

Cross-lamination may be placed in three general types, which are sand-bar, ripple-mark, and dune.

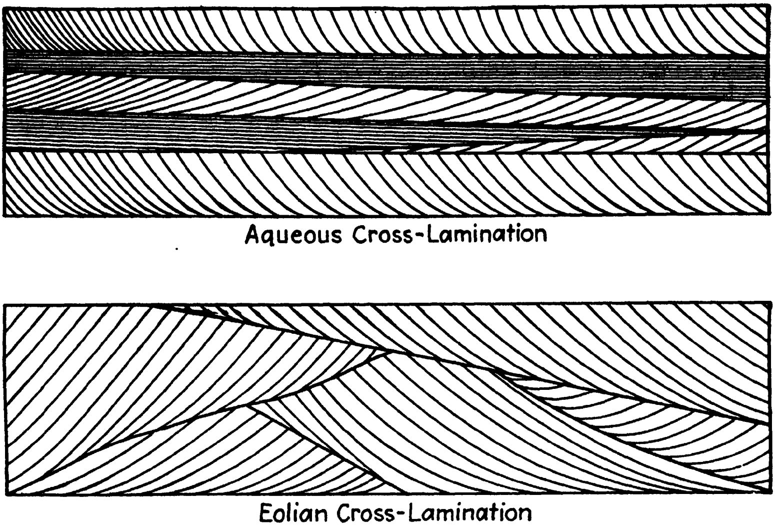

FIG. 62. Diagrams illustrating aqueous and aeolian cross-lamination. The angles of inclination are exaggerated. The wedge-shaped units of aeolian cross-lamination are in striking contrast to the units of aqueous cross-lamination. The smooth phase of lamination is also shown in the upper diagram. Under some conditions aqueous cross-lamination may assume the form of wedge-shaped units, but the scale would be small.

Sand-bar Cross-lamination

Sand bars move over bottoms as plateaulike areas with more or less steep slopes on advancing sides. Heights above the bottom may range from 2 or 3 cm. to 3 meters or more, and foresets may be as great as 30 meters long. Sand, ordinarily derived from erosion of the top and producing truncation of lamination, rolls across the top and the down-current side of a bar. This down-current slope may be straight and normal to the current but is more likely to be irregular, with parts concave and others convex. Foreset inclinations thus show variations in direction, but all have a strong component in the down-current direction. The top surface approximates horizontality, the base may be a surface of some irregularity. Each bar gives rise to a stratum in which the inclinations of the foresets have the same general direction of inclination. Succeeding strata may have the inclinations of foresets in the same or different directions (Fig. 62) depending on whether deposition was in a stream or in standing water.

Ripple-mark Cross-lamination

Wave and current ripple mark produce cross-lamination of different kinds (Gilbert). Current ripple mark has cross-lamination on the lee or down-current side of each ripple, and foresets may have a maximum length of about 15 cm. The inclinations are in the directions of ripple advance, which in streams is mostly downstream and in standing waters may be in several directions. The laminations are truncated at the top, as in sand bars, except in those cases where the supply of sand is so great that ripples are covered as soon as formed, under which condition there is no truncation, the laminae parallel the surfaces of the ripples, and a series of small asymmetrical anticlines and synclines is simulated with axial planes in the synclines inclined up-current and shown by reason of larger particles in the synclines (Spurr). Observed from a distance the effect may be that of pronounced up-current cross-lamination (McKee). This may be more apparent than the actual cross-lamination (Fig. 63). Superimposed cross-

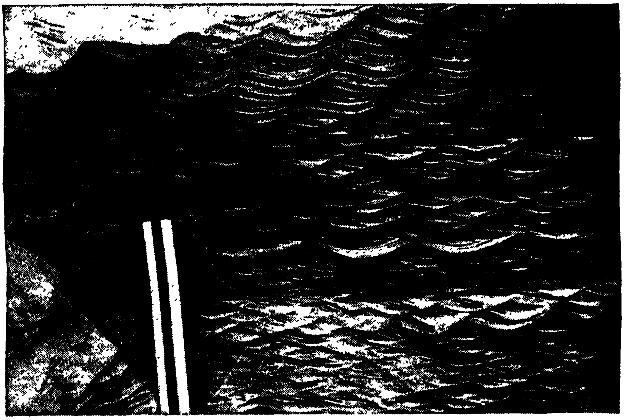

FIG. 63. Ripple laminae showing lee-side concentration of sediment with production of pseudo-cross-lamination in opposite direction. The true cross-lamination is inclined to the right and the currents moved in that direction. The pseudo-cross-lamination is inclined to the left. The latter may be more apparent from a distance. At 123.5 Mile Canyon, Grand Canyon, Arizona. (*Photograph by E. D. McKee.*)

laminated units produce a similar effect (Fig. 67), but the cause is different (McKee, 1939).

Wave ripple mark has laminae meeting at sharp angles beneath the crests. Laminae are short and incline away from crests.

Dune Cross-lamination

Cross-lamination in dunes is inclined in the directions of dune advance, and sand is either derived from windward sides of dunes through erosion, or introduced from other sources. Inclination of laminations is controlled by character and quantity of sands and rates of deposition. The considerable range in direction of winds over short periods may produce much variation in direction of inclination in the different cross-laminated units. There may also be considerable variation in direction of inclination of foresets, as the lee sides of dunes are frequently trough or crescent-shaped in the directions of dune advance; but as pointed out by Beadnell, Shotten, and Reiche, the general direction of inclination of the cross-laminations should be that of the prevailing winds. Foresets tend to be long and, as already noted, the cross-laminated units are wedge-shaped if seen in two dimensions.

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UNCONFORMITY

An *unconformity* is a surface of erosion, or nondeposition, separating two groups of strata. There are two classes, termed *nonconformity* and *disconformity.* In a nonconformity the older or underlying group of strata was deformed and eroded before deposition of the younger or overlying group. In a disconformity there was erosion or nondeposition, but not deformation. There may have been warping. Every unconformity represents absence at the place of a part of the geologic column of which the time value is at least that of a formation. As no one has a very clear concept of the extent of a formation, this yardstick is not a very satisfactory one. A term in some use is *diastem,* which was proposed to apply to a disconformity of less value than a formation, also not a very satisfactory measure. *Ravinement* is a variety of disconformity made in combined delta and shallow-water marine deposits by advance of the sea from any cause over the area of the delta.

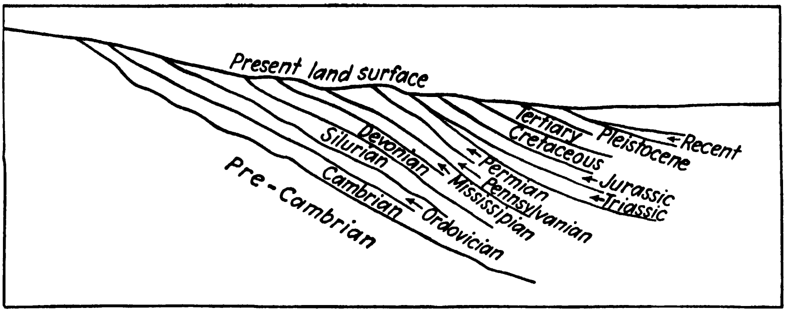

FIG. 64. Diagram showing relations of unconformities to each other and the potential unconformity of the present land surface. Only major unconformities are represented. Had no erosion taken place many of the unconformities would have united as do the Triassic and Jurassic before joining the present land surface. Minor unconformities unite with the major before joining the present land surface.

Each major unconformity either directly, or indirectly through other unconformities, merges into the present land surface, and a composite drawing of all the unconformities in the column would show the land surface over the oldest known rocks extending laterally and repeatedly branching as it spread into the column to coincide with every major unconformity (Fig. 64). It is true that in many places erosion has severed the connections between branches so that doubt may exist as to where unions were located. Every uncomformity has decreasing time value in the basinward direction and ultimately disappears.

Estimates of time values of unconformities are difficult, and particularly difficult is the differentiation of a diastem and a disconformity so far as general appearances are concerned. The relief of an unconformity has little or no significance in terms of time. Great relief is impressive, but an unconformity of great relief may represent only a fraction of the time of one of little relief, and the opposite may also be true. An unconformity separating the oldest Pre-Cambrian from the latest Pleistocene may have the same physical appearance as one between the latest Pleistocene and the middle Pleistocene. The fossils of the strata bounding an unconformity are the only indicators of time value, and these are not always decisive for determinations within narrow limits. A nonconformity may represent a longer time than a disconformity, as the event of deformation is involved, but it by no means follows that such is invariably the case.

Nonconformities are not difficult to recognize, but disconformities may be extremely difficult to detect, particularly if there is little or no relief. There are probably *many in the geologic column still unrecognized* and also probably some that are *considered recognized that are not there.* If there is evidence of erosion the task is not difficult, but if such evidence is wanting and the bounding strata appear parallel, fossils serve as the only evidence, and these may not be definite. Layers of conglomerate *suggest but do not prove* a disconformity, and a layer of phosphatic nodules has the same significance. If the faunas of the bounding formations are elsewhere separated by strata with distinctive faunas, a disconformity may be assumed, but it should not be considered proved. As clastics with some degree of coarseness may be the final sediments deposited before a disconformity developed, and as the initial deposits of the succeeding formation may also be clastics, it is obvious that the disconformity lies in the midst of the clastics. The overlying clastics may have derived some materials from the underlying, of which some may be fossils, and small organisms living when the overlying clastics were deposited may have settled in the interstices of the underlying sediments. The exact position of the unconformity under these conditions is extremely difficult if not impossible of placement.

CONTINENTAL UNCONFORMITIES

Continental unconformities are defined as those of which the surface of unconformity was cut by a continental agent and succeeding deposits are of continental origin. The agent of deposition may have been aeolian, glacial, or aqueous. Glaciofluvial agents of deposition are included in the glacial.

Winds deposit sands over surfaces not previously affected. These are buried and become unconformities. They may have relief from great to none. The time value may be little or much, and the unconformity may be of either kind. Winds frequently erode their earlier deposits and later may bury the eroded surfaces. Owing to possible marked divergence of dips of the cross-laminations, an appearance of nonconformity may be presented. The time value is likely to be very short.

Each advance of a glacier leads to some erosion of the surface over which advance was made and deposition over the surface in advance of the glacier, transforming the buried surface into an unconformity. As a glacier retreats, it buries the surface released beneath glacial and fluvioglacial deposits, and the buried surface becomes an unconformity that may be of either kind. The time interval may be short or long.

Fluvial unconformities may be placed in two classes, of which one is related to alluvial fans and deltas and the other to the valley-flat part of a stream.

Alluvial-fan and kindred deposits are deposited on surfaces of varied origins over which they extend outward from the uplands and also rise on them. The surface buried becomes an unconformity that may represent a very long or a very short time and may be either a nonconformity or disconformity.

Unconformities in undisturbed delta deposits may be either marine or continental and are invariably of the disconformable variety except perhaps at the base, where a delta may be built over deformed strata. A delta advances or retreats depending on the balance between supply and disposition of sediments. If the disposition ability exceeds the supply, the delta front is eroded and retreats, leaving behind an eroded surface. A change in conditions may cause supply to exceed disposition ability, and the delta front advances. A disconformity separates the two deposits, which may be fluvial, paludal, lacustrine, or marine deposits below and fluvial deposits above. The time interval is likely to be short. If sea level rises, the subaqueous deposits are eroded unless submerged below the base level of erosion, and even then some erosion is likely as the sea advances over the surface submerged. Advance of the delta buries the eroded surface, and it becomes an unconformity of the *same character as the preceding but formed by different conditions.* This has not always been appreciated.

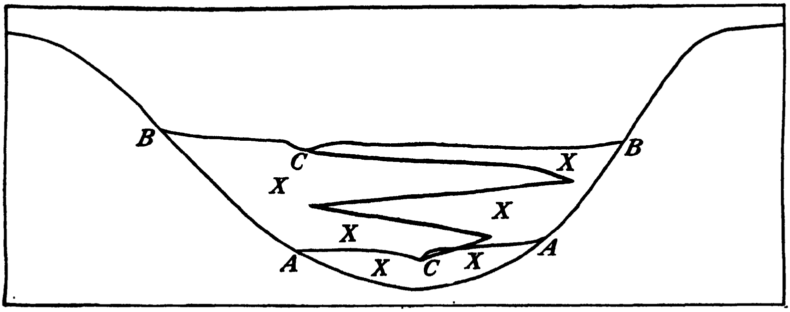

Fig. 65. Trace of an erosion surface—an unconformity—in the floodplain deposits of an aggrading stream. *A-A*, initial stage. *B-B*, level of maximum aggradation. *C-C*, trace of migration of a meandering stream as deposits accumulated. *X*, fluvial deposits. A well drilled in these deposits might cross the same unconformity four times.

Unconformities in the valley flat arise in the channel deposits owing to the phenomenon of cut-and-fill. Unconformities are extremely abundant, local relief may be great, and correlations of the unconformities are difficult if not impossible. The relations are those of disconformity, but as initial dips may be high, a pseudononconformable relationship may exist.

An aggrading and meandering stream migrates laterally and develops a surface of erosion that is progressively more or less deeply buried by fluvial deposits. This produces a peculiar type of disconformity (Fig. 65). It tends to have considerable relief and rises in zigzag manner in the deposits. It is the same surface of erosion and the same disconformity. The time interval may be short or moderately long.

MARINE UNCONFORMITIES

Marine unconformities are of great variety but may be placed in four classes:

1. Rapid rise of sea level progressively submerges a land, producing what is known as progressive overlap. If rise is very rapid, there may not be much erosion. The coarse debris of the land area forms a basal conglomerate, and there may be buried soils. The relief of the unconformity approximates that of the submerged surface. As deposits are made in the advancing sea, each depth is characterized by sediments related to depth, and as depth at a place changes with advance, the sediments and faunas related to depth and character of bottom move landward with the shore. Thus the bedding *has one inclination, but sediments and faunas related to depth have another, and thus faunas and lithology diagonally transect the time units.* If organisms are not closely adapted to depth and bottom conditions, or depths are moderate and dead organic matter can be spread by waves and currents, paleontologic units may coincide with time units, and lithic units may transect both time and paleontologic divisions (Fig. 66).

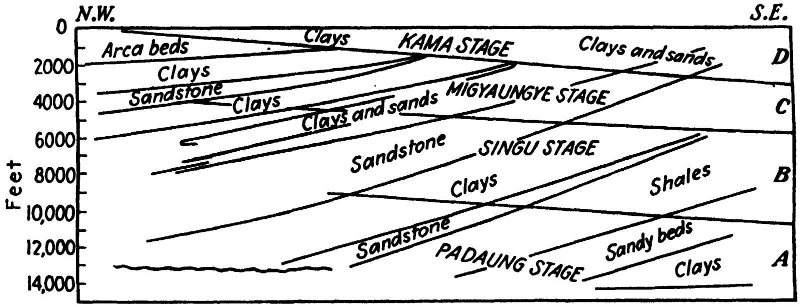

FIG. 66. Diagram showing about 30 miles of the Pegu System of the Tertiary of Burma. The section is drawn from northwest to the southeast, the lithic units rising to the southeast, the faunal units to the northwest. Lack of coincidence of lithic and faunal units is indicated, the two units showing a diachronous behavior. The paleontologic divisions transect the lithic divisions. (See Fig. 2.) [*After L. D. Stamp, Geol. Mag.*, 62 (1925), 520.]

2. Slow rise of sea level leads to development of a wave-cut surface that is usually on solid rock and may ultimately extend for long distances from shore. Sediments from the shore must be carried across this surface, and a considerable interval may intervene between detachment from parent rock and final deposition. This is also progressive overlap but different from the preceding. As the materials from the shore are subject to much movement in the shore zone before transportation out to sea, it

may be expected that the sediments finally deposited are composed of fine particles. Rise of sea level ultimately brings the lower part of a wave-cut surface below the base level of erosion and thus permits sediments to be deposited upon it. These may be fine-grained clastics or organic. A basal conglomerate is not probable. Submergence of an area in the old stage of the cycle of fluvial erosion produces an unconformity on a wave-cut surface of little relief, but submergence of an area in an earlier stage of the cycle yields relief dependent upon the rate of rise of sea level, the ability of the waves to erode, and the relief of the original surface. The unconformity may be of either kind.

3. Rise of sea level interrupted by periods of stability may permit bottoms to be built to the base level of deposition for considerable distances from shore during each period of stability. The unconformity at the base of the deposits has diastems branching from it into the deposits, each for a period of stability and each marked by much washed, more or less coarse bottom materials, the residue of sediments from which the fine particles were transported to deeper waters. If periods of stability should last sufficiently long, some removal of previously deposited sediments is possible.

4. Falls of sea level lead to lowering of the base level of erosion, removal of previously deposited sediments, and creation of an eroded surface that becomes an unconformity should new conditions permit deposition upon it. Increase in the quantity of materials brought to a sea bottom may cover this surface without further movement of sea level.

Stratigraphic breaks and unconformities may be indicated in coral and other reefs by lateral expansions of the reefs and reef debris during the times of stability as upward growth ceases when sea level is reached and lateral expansion is possible. Rise of sea level again permits upward growth and discourages lateral growth. A disconformity is near the top of the lateral growth.

SUMMARY

Unconformities in continental deposits may have considerable relief and limited extent. They are likely to be marked by coarse deposits; those of marine deposits are likely to have limited relief and considerable extent. The time value of an unconformity is not measured by its relief, and fossils in most cases provide the only fairly reliable measure of the time involved. Marine unconformities may or may not be indicated by clastics, and these may be the top of the underlying strata, bottom of the overlying strata, or parts of both groups of strata. Clastic sediments in the midst of nonclastic sediments do not prove an unconformity. It is probable that there are more unconformities without the traditional basal conglomerate than with them. Disconformities may be extremely difficult of determination, and it is probable that many have not yet been recognized. Times of great stability may build up many parts of shallow bottoms to base levels of deposition, which may become unconformities if there is elevation of the sources of supply or rise of sea level. Emergence followed by erosion and subsequent submergence is not essential for development of a marine unconformity.

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RIPPLE MARK

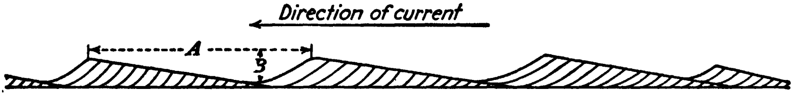
INTRODUCTION

Ripple mark is made on surfaces of little-cohesive materials by currents of air and water. Ripple mark made by water is termed aqueous, that made by wind, aeolian. Aqueous ripple mark is made only on granular sediments, and its presence in any rock plainly indicates the granular character of the composing materials at the time of final deposition. Aqueous ripple mark is never made on muds, and if a deposit of sand passes abruptly into one composed of mud, ripple marks end abruptly where the mud begins. Air currents may make ripple mark on dusts as well as sand. Aqueous ripple mark may be made by waves and currents; aeolian ripple mark is made only by currents.

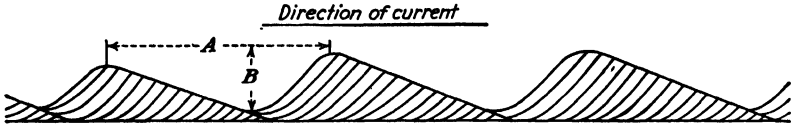
Consideration of ripple mark requires definition of several terms (Kindle): *amplitude* is defined as the elevation of the crest of a ripple above the adjacent troughs; *wave length* is the distance between adjacent crests or troughs; *ripple index* is the ratio of wave length to amplitude; a term applicable to current ripple mark is the *horizontal form index*, which is defined as the ratio of the length of the up-current to the down-current side.

WAVE RIPPLE MARK

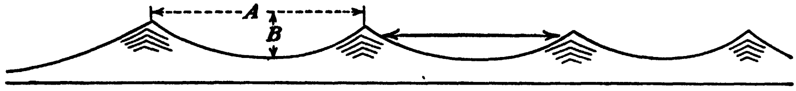
Wave ripple mark, also termed oscillatory and symmetrical ripple mark, is characterized by symmetrical slopes, sharp crests, and rounded troughs (Fig. 67). Double and triple crests develop in some cases. The mechanism of formation has been stated by Gilbert. Water particles in wave motion move in more or less circular orbits at the surface of the water. This motion is communicated downward, but the circular orbits change to elliptical with the longer axes horizontal. Close to the bottom the ellipses are nearly flat and the motion is essentially a swinging back and forth. Any prominence on the bottom determines



Air Current Ripple Mark and Cross-Lamination



Water Current Ripple Mark and Cross-Lamination



Wave Ripple Mark and Cross-Lamination

FIG. 67. Diagrams of current (asymmetrical) and wave (symmetrical, oscillation) ripple marks. The diagrams also show the cross-lamination developed in formation. In current ripple mark the back slope is inclined up-current, resulting in a pseudo-cross-lamination in that direction when different cross-laminated units are superimposed. *A* is the wave length, *B* the amplitude. Wave length divided by the amplitude equals ripple index. It is obvious that the ripple index of the wind ripple mark will be greater than the others.

vortices alternately on the two sides, and ultimately a ridge develops with symmetrical slopes and a sharp apex. Rearrangement of the ridges takes place as they develop. Amplitudes and wave lengths are functions of wave size, wave period, and depth of water.

Wave ripples once formed are approximately stationary and do not advance in either direction. Crests and troughs hold constant positions as long as producing conditions do not increase in magnitude. Decreases in magnitude do not affect them. After formation they have a surprisingly high degree of firmness.

Wave ripple mark is formed on all sand bottoms of which overlying waters are agitated by waves that move the sands, and they may be seen on the shallow bottoms of most ponds, lakes, and seas whose waters are not too strongly moved by currents.

Burial and preservation of wave ripple mark are not difficult, if covering sediments settle from suspension, but are difficult if bottom currents bring the sediments, as such currents destroy the ripples in a very short time in spite of the firmness noted above. Rapid burial seems essential for preservation.

Since diminishing wave strength does not destroy wave ripple mark, it may result in the formation of other ripples across earlier ones if weaker waves stir the overlying waters.

Wave ripple mark may form to considerable depths. Waves are reported to stir the bottom in the English Channel to the depth of about 75 meters, to 100 meters on the banks of Newfoundland, and to 200 meters in the open ocean. Chafing of cables by waves to the depth of 480 meters has been reported. Wave ripple mark has been reported near the island of Réunion, (Bourbon) in the Indian Ocean at the depth of 188 meters, but it is not certain that the ripple mark was of wave origin.

Some relation exists between dimensions of wave ripple mark and the depth of water in which it forms, but not a great deal of information is available. It is known that wave length increases from the surface downward to some limiting depth and then decreases in dimension to final disappearance. Gilbert observed that in Lake Ontario "for moderate depths the size of ripples is not very sensitive to depth." Since large waves cannot form in very shallow water, it is obvious that wave ripple mark formed there should have correspondingly small wave length and amplitude, and Kindle's observations show that waters less than 15 cm. deep have wave ripple mark with amplitudes of 5 mm. or less. But ripple mark of that dimension may also be found in waters of all depths to the deepest in which ripple mark has been determined. Small wave ripple mark is thus not an indicator of depth, and ripple mark of large wave length and amplitude indicates neither very deep nor very shallow water, but water of some intermediate depth. The wave length of wave ripple mark ranges from about 0.5 to 50 cm., with most in the range from 3 to 12 cm. Amplitudes range from 0.1 to 4.5 cm., with most in the range from 0.5 to 1.5 cm. The ripple index range is from about 4 to 10.

Wave ripple mark is valuable for determining the attitude of strata, as the sharp crests project outward on top of beds and the rounded troughs are indented. Counterparts made on bottoms of immediately overlying strata have the crests and troughs in reverse positions. The top of a bed is thus readily determined. This may serve to solve difficult structure.

AQUEOUS CURRENT RIPPLE MARK

Aqueous current ripple mark (Fig. 67) differs from wave ripple mark in that slopes are asymmetrical and crests are rounded. Currents are responsible for formation. Below some velocity that varies with depth and character of sands, a current is not able to move the sands over which it flows. As velocity is increased, some particles begin to roll. This is the first critical point. It varies with depth of water and character of sands. Fine sands of 0.4 mm. mean diameter at depth of 0.13 meter require a velocity of 0.26 meter per second, and gravel of mean diameter of 7 mm. at depth of 0.066 meter requires a velocity of 0.86 meter per second to start motion. After the first critical point is attained, ripple mark begins to form as numerous more or less equidistant ridges trending at right angles to the current. These ridges have gentle slopes on the stoss or up-current sides and steeper lee slopes. Particles are rolled up the stoss side to the crest and then roll down the lee slope. There is a vortex on the lee side that causes small particles to roll up lee slopes for short distances.

Current ripple marks migrate down-current, although the ripple at any time is rigid and only the top particles of the stoss slope are in motion. The steep slope of current ripple mark is down-current, which in streams ordinarily means downstream but not necessarily down-valley as meanders may lead streams for short distances up-valley. The considerable variation in direction of currents in standing bodies of water produces ripple mark with lee slopes inclined in several directions. Lamination in current ripple mark parallels lee slopes and has inclinations determined by velocity of currents and supply of sands, but it is thought never to exceed 43 deg. and is commonly less than 30 deg. The sands in current ripple mark are coarsest in the troughs and finest on the crests, and, if the supply of sand is sufficiently great, current ripple marks are buried as formed. Crests and troughs of successive ripples then overlie each other, there is lamination on both stoss and lee sides, and the strata have the appearance of being asymmetrically crumpled. Seen from a distance the coincidence of crests and troughs may produce apparent cross-lamination in the up-current directions (Fig. 63), as the sands are of coarser grain in the troughs than on the crests.

Increase of velocity of a current brings it to the second critical point, and current ripple mark ceases to form. The force of the current extends below the bottom so that the sands are lifted and flow with the water. Further increase in velocity brings a current to the third critical point. The bottom then again becomes waved but different from that between the first and second critical points in that the ripple marks travel upstream through erosion on the down-stream and deposition on the up-stream side. Crests are rounded and profiles are more or less symmetrical. Such lamination as exists is inclined up-current and is made on the up-current slope. This form of ripple mark was designated antidune by Gilbert and *regressive sand* waves by Bucher. Regressive sand waves have but momentary existence. There is little chance of preservation, but they have been reported in Ordovician sandstone of Ireland.

Water currents of high velocity above the third critical point and transporting large volumes of sands develop large ripplelike structures that have been termed *progressive sand waves.* These are large irregularly shaped ridges with orientation transverse to the direction of the forming currents. Crests are broad and gently rounded, and slopes are often nearly symmetrical. Movement seems to be taking place throughout the interior of the wave, greatest at the summit and decreasing downward. Movement of these waves is down-current. Reduction of velocity below the third critical point causes them to become rigid, a vortex forms on each lee side, sand grains roll up stoss slopes and drop over the lee slopes, and they assume the profile of a current ripple. These are the *meta-ripples* of Bucher. They have distinctly asymmetrical slopes with inclination on the up-current side of 4 to 6 deg., and 14 to 33 deg. on the down-current side. Wave lengths range between 90 and 1,800 cm. and heights between 10 and 70 cm., with most between 20 and 40 cm., although they may have heights exceeding 2 meters. Comparison of ripple indices shows that increase in wave length is not proportional to height (Johnson).

Structures comparable to meta-ripples are the *fulls and lows* of British coasts. These are uniformly rounded, low sand ridges situated in or just below the intertidal zone of some coasts. The ridges trend parallel to the beach and may rise 1 meter or more above the separating troughs. The ridges are the *fulls* and the troughs the *lows*. Formation seems to be connected with breakers and the currents thus produced. Evans has described similar features which form in the shallow waters off the eastern shores of Lake Michigan under the name of *low and ball*. He states they are formed by breakers.

Many limestones have large ripplelike structures that vary from very symmetrical to very asymmetrical. These have been designated *para-ripples* by Bucher. They resemble meta-ripples, from which they differ chiefly in the greater abundance of symmetrical slopes and in a smaller index in the symmetrical ripples. They are ascribed to current action chiefly on the basis of poor sorting of the composing materials. The origin is not known, but it may be that they are meta-ripples that have become modified by wave action.

Aqueous current ripples and their counterparts have like profiles, and it is impossible to determine the original from the counterpart on general appearance. Cross-lamination related to the ripple formation, however, is only in the original, and this forms on the top of a bed. Original position of a ripple-marked bed is thus determined.

Current ripple mark may be made at all depths to which currents extend. These are known on the coast of Norway to the depth of 250 meters, to about 300 meters on the coast of New England, to 800 meters about the Azores, to 2,000 meters in a channel between some of the Canary Islands, and to 1,500 meters between some of the islands of the East Indies. Current ripple mark is thus not indicative of any particular depth except that large ripple mark can not form in very shallow water. Wave lengths of simple current ripple mark range from about 1 to 25 cm., amplitudes from 0.5 to 5 cm., with most in the range from 0.3 to 1 cm. Indices range from 4 to 20, with most from 4 to 10. The horizontal form index ranges from about 2 to 4. Wave lengths increase with current velocities.

Complex Forms of Aqueous Ripple Mark

Currents are rarely simple, and much ripple mark represents action of complex currents or the combined action of two or more currents. These may be differentiated into several types.

LINGUOID RIPPLE MARK

Current action seems to be largely if not wholly responsible for ripple mark with tonguelike projections and on this account termed linguoid. The pattern is highly irregular, but the tonguelike outline is distinctive. A current under many conditions seems to be composed of different streams of which some are stronger than others. The former advance parts of a ripple ridge and thus produce tonguelike outlines. This form of ripple mark is best developed on bottoms of shallow streams.

RHOMBOID RIPPLE MARK

The rhomboid form of ripple mark is extremely common on sand beaches of fairly gentle to moderate slope, where it forms during the final stages of backwash of each retreating wave. A certain velocity seems essential for formation, and it does not seem to form on very steep beaches. Woodford states that relatively steep slopes are required, which are "perhaps 5 deg. or 10 deg. or 15 deg." The writer has seen them on slopes of 3 to 4 deg. The pattern resembles the scales of a ganoid fish or surface of a Lepidodendron tree. The mark has rhomboidal or diamond shape, of which the acute angles are normal to the beach and parallel to the currents of the retreating waves. The sides of the down-current angle are bordered on the outside by steep lee slopes, which in turn are inclined into the upper halves of the bordering rhomboids (Fig. 68). Each rhomboid ranges from 12 to 25 mm. wide and from 25 to 50 mm. long. The short axis is always parallel to the shore. The slopes that define the rhomboids are not more than 1 mm. high, and very small changes of current suffice to destroy them. After the sands composing the rhomboids become dry the latter are very quickly destroyed by winds of low velocity. There is little chance of rhomboidal ripple mark being preserved, but they have been recorded in the column (Bethune).

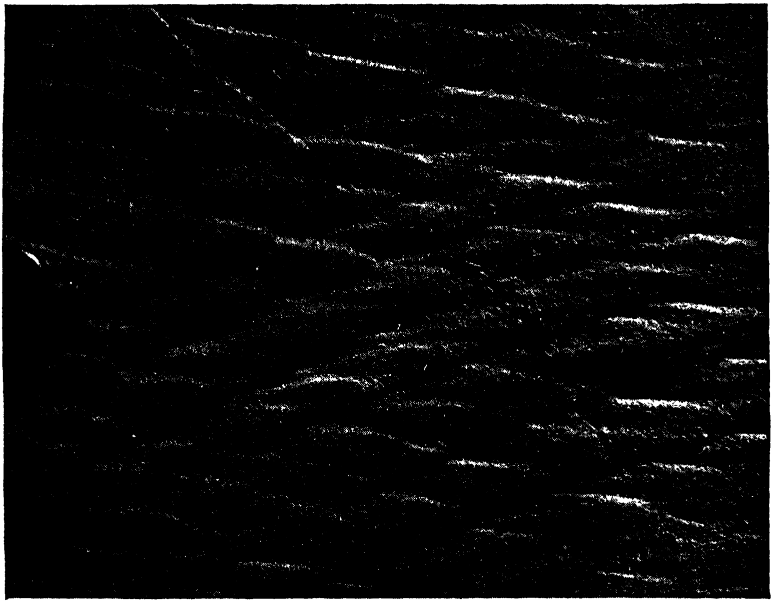


FIG. 68. Rhomboid ripple mark (photograph of a plaster cast). Mark made as a wave rolls down a beach. The rhomboids are depressions above, and plateaus below. It may seem that the reverse is the case if the illustration is observed from certain angles. The watch is on the upper side of the beach. [*Taken from E. M. Kindle, Recent and fossil ripplemark, Geol. Surv. Canada, Mus. Bull.* 25 (1917), *Pl.* 19*b*.]

OSCILLATION CROSS RIPPLE MARK

Two sets of waves acting concurrently or successively or a set of waves acting on current ripple mark produce wave interference or cross ripples. These may be rectangular or hexagonal. Rectangular cross ripple mark consists of two sets of ripple ridges intersecting at right angles and enclosing rectangular pits. In the hexagonal variety, parallel ridges are arranged in zigzag fashion with obtuse angles in adjoining ridges facing in opposite directions. Crossbars connect apexes on opposite sides of each ridge, and the enclosed pits tend to be bounded by six sides. The rectangular ripple marks are made by simultaneous or successive waves, or waves successive to currents, that oscillate at 90 deg. to the direction of the ripple marks first formed. The hexagonal cross ripple marks seem to be made by waves

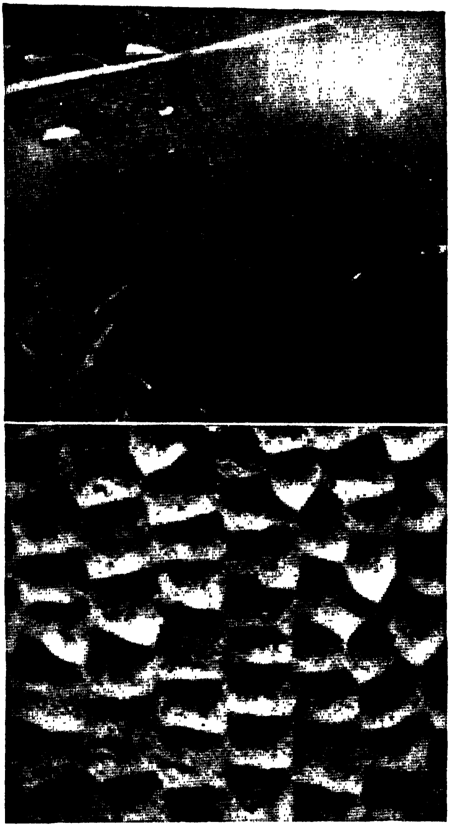


FIG. 69. Supposed "tadpole" nests (top); interference ripples (bottom). (*Photograph by E. M. Kindle.*)

acting under the same conditions but at some angle between 90 and 45 deg. The two varieties may be seen on the same slab, and pure forms of either are not common. After several days' existence the pits of both kinds of ripple mark tend to become round and the ridges to assume circular outlines. Wave interference ripple mark is common on the bottoms of ponds formed between dumps in the strip coal-mining region of southwestern Indiana. At one time the small circular depressions were considered to have been made by tadpoles, and two species were differentiated and given scientific names (Fig. 69).

CURRENT CROSS RIPPLES

A set of current ripples may have another set developed upon it by a later current moving in a direction different from that which made the first set. The second current should be of less strength than the first and should not last for a long time, or the first set is destroyed. Current cross ripples are the result. They may intersect at any angle.

Ripple Mark of Combined Wave and Current Origin

Ripple mark of either wave or current origin may be reshaped by the other agent. A bottom may be agitated by a current moving in one direction and the waves in another. The ripple mark formed depends upon the strengths of the waves and currents. If the currents are strong and the waves weak, the wave ripple marks form on the gentle slopes of the current ripples and are of small size. If the waves are strong and the currents weak, the wave ripples are the larger and the surface is knobby. Wave ripples developed under these conditions are rarely symmetrical. Current ripples may be totally destroyed by wave action as fast as they begin to form, and only wave ripples remain. The opposite may also take place.

Aqueous Ripple Mark and Composition of Sediments

Aqueous ripple mark is formed only in granular sediments, but the granules may have any composition. As most sands are composed of quartz, it results that most ripple marks are on sands of this composition and are preserved on quartz sandstones. But many beaches and adjacent bottoms are covered with sands of calcite, aragonite, or dolomite. These sands contain ripple marks of kinds and dimensions similar to those on quartz sands under similar conditions. Many limestones of the geologic column have ripplelike structures that are very large with wave lengths ranging from 0.33 to 2 meters. They range from slightly asymmetrical to nearly symmetrical, and the crests are rounded. They are obviously not complete products of wave action and are evidently a complex form in which both waves and currents have participated in forming.

Tabular Arrangement of Aqueous Ripple Mark

The various varieties of aqueous ripples and their origins arranged in tabular form are as follows:

I. Wave motion.	Wave or oscillation ripples, symmetrical.
II. Simple currents.	
1. Current velocity between first and second critical points and no granular sediments in suspension.	Current ripples, asymmetrical. Persist after current subsides. *a.* Normal. *b.* Rhomboid. *c.* Linguoid.
2. Current velocity at or above third critical point with much granular sediment in suspension.	Sand waves, symmetrical, exit only while current lasts. *a.* Regressive sand waves, at lower velocities. *b.* Progressive sand waves, at higher velocities.
3. Falling velocities of high range causing transformation of unstable sand waves.	*a.* Meta-ripples, asymmetrical. *b.* Pararipples, symmetrical and asymmetrical.
III. Currents, or currents and waves, of different direction acting simultaneously or in succession. Cross ripples.	1. One set forming after completion of other. *a.* First set current, second set wave, wave or oscillation cross ripples. *b.* Both sets current ripples, current cross ripples. 2. Two sets forming simultaneously. Compound ripples.

AEOLIAN RIPPLE MARK

Ripple mark made by wind is always of the current variety and always asymmetrical with the windward or stoss slope of little inclination and the leeward slope steep (Fig. 66). Ripples develop in any loosely coherent materials, and they form in dust and snow. However, ripples in snow, probably connected with the tendency of snow granules to cohere, tend to be less steep on windward than leeward sides. Most aeolian ripples are of the dune type, and only a single case of antidune ripple mark of aeolian origin has been recorded.

Wave length and amplitude of aeolian ripples seem to vary little with wind velocity, but they do vary directly with dimensions of particles, and the largest wave lengths and amplitudes are in the coarsest sands. Most wave lengths are between 5 and 10 cm. with variations beyond these limits; most amplitudes are 2 to 4 mm. Most ripple indices range between 20 and 50. Wave lengths as great as 25 cm. have been observed in coarse sands. Aeolian ripples approach parallelism but do anastomose in broad netlike patterns. Aeolian ripple marks seem rarely to have been observed in the geologic column, but McKee (1934) states occurrence in the Coconino sandstone (Permian) of Arizona.

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CONTEMPORANEOUS DEFORMATION OF SEDIMENTS

Introduction

Within recent years there has been a progressively increasing appreciation of the importance of the deformation of soft sediments by forces not connected with diastrophism, and the danger at present is that the pendulum may swing too far and some deformation due to diastrophism may be included. Criteria for differentiation of structures due to diastrophism from those connected with sedimentational processes have not been clearly established, and in many cases personal preferences seem to determine assignment. Four more or less distinct surface causes seem to be responsible for most soft-rock deformation. In suggested order of importance these are (*a*) sliding (gliding), (*b*) compaction, (*c*) lateral movement through pressure of overlying sediments, and (*d*) thrust of surface agencies. Two additional nondiastrophic causes producing deformation of indurated sediments are recrystallization and removal of materials by solution, mining, or other causes.

Deformation Caused by Sliding

Sliding or gliding takes place in sediments that do not have adequate support. This may be due to deposition on too steeply inclined surfaces, excessive local deposition creating a steep slope, and removal of support by erosion, falling of water level, or melting of ice. Sliding is favored by an abundance of water in sediments, as this adds to weight and increases mobility. Sliding may be subaerial or subaqueous.

Angular sands may rest on slopes of 43 deg. and rounded sands on slopes of 38 deg. but they will not remain there unless deposition is of uniform thickness over an entire slope, and any slight seismic or other shock may start movement. Clays in any quantity can not remain on slopes exceeding 30 deg., and very slight shocks suffice to start movement of clays resting on slopes near the maximum limit. Much sliding of sediments containing clays is known to have taken place on slopes of 10 to 15 deg., and one extensive slide in Lake Zuger, Switzerland, is known to have moved over a slope of 2 deg. 31 min. This slide was 1 to 12 meters thick and 25 meters wide, and it extended into the lake for a distance of 1,200 meters from the shore and to a depth of 45 meters. It seems probable that the greater the volume of sediments sliding, the lower the slope on which movement can take place. Fairbridge suggests slumping as responsible for intraformational disturbances in Carboniferous varved rocks of Australia.

Subaqueous deposition of sediments takes place on surfaces that range in inclination from zero to nearly 90 deg. Steep slopes may be due to erosion of a land surface later submerged, deposition, vulcanism, and diastrophism. Extensive areas of the sea bottoms in the vicinity of the East and West Indies, Japan, the Philippine Islands, and elsewhere have slopes created by diastrophism or vulcanism that are too steep for permanent retention of any sediments.

Rapid local deposition of sediments is not uncommon on the fronts of deltas, particularly those of bodies of water without tides and with weak waves and currents, and deposits are frequently built above positions of equilibrium with consequent later sliding. Sliding is also likely on a considerable scale about coral reefs, as the organic structures are often built into steep slopes on which sediments find temporary resting places but slide later.

The cut and fill of rivers produces slopes on the sides of deeps that encourage sliding. Melt-water deposits built around glaciers are certain to slide when the ice melts. This may often be seen in delta kames of the last ice sheet.

Sliding of marine deposits brings sediments of one depth of bottom upon bottoms of greater depth, increases the thickness and number of beds of sediments at places of landing, decreases the thickness and number of beds at places of starting, places older on younger beds, develops local unconformities, and brings organisms and sediments characteristic of a certain depth to a greater depth. Materials higher on a slope shove upon those below. This leads to the formation of asymmetrical or overturned, more or less closed folds in the materials of the lower part of the slide and asymmetrical, more open folds above. The axial planes of most dip into the slope. Stretching on the axes of the folds is favored, and there may be some faulting. The thickness of the strata involved in a slide, the character of the sediments, the nature of the underlying and bordering materials, and the steepness of the slopes are factors that influence the detailed structures of the deformation. If the sliding sediments are to any degree consolidated, they are very likely to become broken and to form a glide or slide breccia. If large clastic particles are in the sediments, the sliding is likely to place these in positions different from those of deposition, so that they stand at many angles to original bedding planes and form an edgewise conglomerate. After sliding has taken place, subaqueous erosion may level the top of the slide, and thus subsequent sediments hold unconformable relations to those below. The diagram of Fig. 70 ideally illustrates results of sliding.

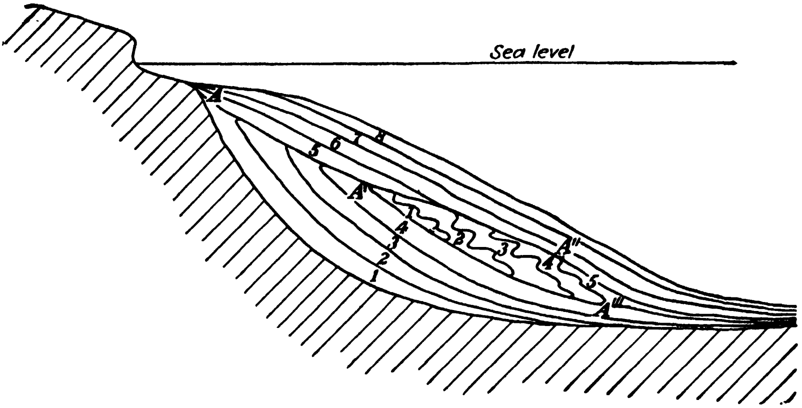

FIG. 70. Diagram to show effects of sliding of sediments. *A-A'*, surface from which the slide was made; *A'-A''*, top of the slide after erosion of the upper part has taken place; *A'-A'''*, surface upon which the slide was made. The section of origin of the slide has been thinned, the section at the place where the slide stopped has been thickened and strata have been duplicated, and there has been displacement of facies. An unconformity separates the sediments that did the sliding from those that overlie.

Deformed sediments referred to subaqueous sliding are present in Miocene lake marls at Oenigen, Germany, the Jurassic of

Solnhofen (the "Krumme Lage"), the Triassic Muschelkalk of the Main region of Germany, the Cape Bon Ami beds of the Devonian of Gaspé Peninsula of Quebec, and limestones associated with reef structures in the Kokomo limestone of northern Indiana, and elsewhere.

DEFORMATION DUE TO DIFFERENTIAL COMPACTION

Deformation caused by compaction arises from settling of sediments over any rigid body. This may be a buried hill, a coral or other reef, or a sand or gravel lens. Compaction is due to expulsion of water, closer packing, and new combinations of the sediments. It may take place to some degree in all kinds of sediments but is generally negligible in shell materials and sands and coarser clastics and is mostly important in fine-grained sediments like clays, silts, and calcareous muds. Marls, clays and some organic sediments in some Wisconsin lakes contain as much as 70 per cent water, some more than 90 per cent, and many fine-grained sediments contain more than 50 per cent. Much of this water ultimately is expelled, with a somewhat corresponding reduction in volume. The compaction due to crystallization and closer packing is also important but is mostly overshadowed by that due to expulsion of water.

Buried coral and other reefs, buried hills, and lenses of shelly sediments, sands, and other coarse clastics undergo little or no compaction, but fine-grained sediments deposited over and around these features may experience compaction to the extent of 50 per cent or more. Strata passing over buried noncompactible materials thus become arched into domes and anticlines, depending on shapes of the buried features.

Deposits on high places or hills on the sea bottom tend to be coarser than those in the deeper places, and more sediments are deposited in the deeper places because of washing of the fine sediments from high places. If the slopes from the high to the low places are sufficiently steep there is subaqueous sliding. The greater deposition over the low places decreases the slope to the higher areas, and, if there was no compaction, the bottom would ultimately be brought to the same level. Compaction and settling, however, proceed simultaneously with deposition, thus maintaining some slope until the high places have been completely buried, and these overlying formations then may become warped through many hundreds and even thousands of meters of sediments. Thus, a rigid body may be reflected through great thicknesses of strata.

These structures have been termed compaction folds. They tend to cover progressively greater areas in higher strata than the rigid bodies to which they are due, inclinations become less with greater distance above the rigid bodies, sedimentary units that surround and pass over them tend to have greater thicknesses outward from the apexes of the structures, and dips of marginal strata increase with depth.

As compaction folds are related to ancient physiography, coral and other reefs, or depositional units of little or not at all compactible sediments, it naturally follows that compaction folds do not follow diastrophic patterns. The writer has termed the trends *amoeboid*. As little-compactible sediments are somewhat erratic in distribution, it may happen that coarse-grained sediments or coral reefs of later times may overlie fine-grained sediments of earlier times that had been deposited in basins. Thus, compaction folds may rise in higher strata above basins in lower strata, and basins in higher strata above compaction folds in lower strata. That is, the compaction folds of one horizon need not be at the same places as compaction folds in overlying or underlying horizons.

Structures of compaction origin can not fail to be common near bases of formations that rest upon a buried mature-to-youthful land surface or in the sediments of shallow water where reef structures form and lenses of clastics and shell matter are deposited in the midst of muds. They have been noted in the Mid-Continent and other regions in the drilling for oil and have been seen in section where coral reefs are exposed.

DEFORMATION DUE TO FLOW OF SEDIMENTS CAUSED BY OVERLOADING

The deposition of sediments places a load upon those already deposited, and if the load is differential, as is probable, fine-grained sediments move outward from beneath the greatest load. This is readily demonstrated in the laboratory, and experiments have shown that fine-grained sediments move toward places of little load with more or less crumpling as a consequence. Coral reefs built upon soft sediments may settle through flow of underlying sediments, with the result that underlying sediments incline toward the reefs. This has been seen on the Island of Gotland and elsewhere. The under dipping beds are in strong contrast to the higher strata where inclinations are away from the reefs. The salt domes are built of evaporites which have flowed from beneath loads to places of less load, and there may have been much development of intricate structure in connection with the flow. The structures made in this way in evaporite deposits, coupled with deformation resulting from recrystallization, have been termed *enterolithic*, from resemblance to the intestines of animals.

Rettger's experiments showed that differential loading produced overturned folds with axial planes inclined toward the load and trends parallel to the edge of the load. Both normal and thrust faults were developed, which were downthrown on the side of the lesser load.

DEFORMATION FROM THRUST OF SURFACE AGENCIES

Considerable deformation is known to have been caused by thrust of glacial and floe ice and icebergs and sliding masses of earth and rock. Glaciers pushing against stratified deposits, particularly those made by melt waters, produce deformation. The deformation is expressed in brecciation, thought to be generally near the ice, overturned folds and faults beyond the brecciation, and more gentle folds with distance. The complexity of the deformation may be extremely great. Examples have been described from the Titt Hills of Alberta, Martha's Vineyard and Block Island of the New England Coast, England, Denmark, and elsewhere (Fig. 71).

The thrust of floating sea ice at the present time is probably deforming sediments of all bottoms with which the ice comes in contact. The deformation naturally affects only certain layers, which may be one or many. It consists of thrust faults and folds that range from open to overturned. The deformation usually has limited area for any one level. Pleistocene glacial lake clays locally show, in wonderful perfection, deformation produced by floating ice, and it has also been observed in lake deposits of past glaciations (Fig. 72).

Sliding masses of earth and rocks may deform stratified materials over and against which they move, and every such slide into bodies of water may have deformed sediments in advance of the slide.

Some limited deformation is produced by the thrust of lake ice on shores and on sediments of shallow bottoms. This is locally expressed in folding and faulting, particularly where the thrust is against materials of swamps.

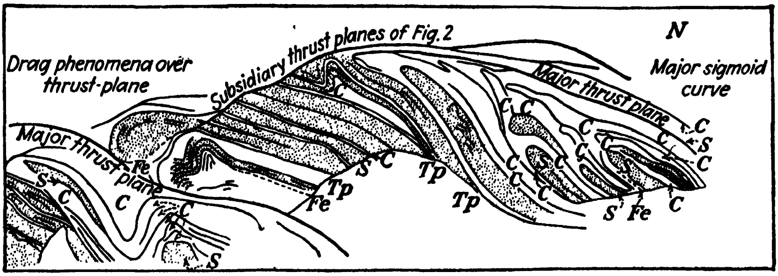

FIG. 71. Deformation due to thrust of glacier. Partial section across the Mud Buttes of Alberta showing 350 to 400 ft. The thrust was from the right. Dotted areas are sands, unmarked clays. [*After G. Slater, Bull. Geol. Soc. Am.*, 38 (1917), 724.]

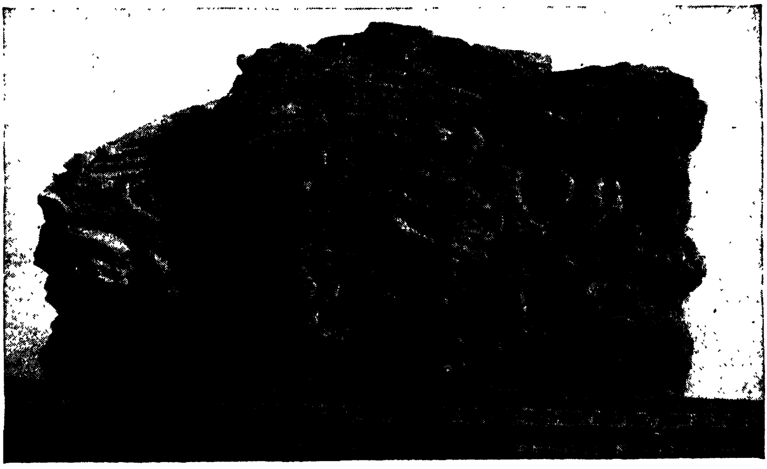

FIG. 72. Deformed varves in the Waupaca shale pit, Waupaca, Wisconsin. The deformation is referred to shove of floating ice. The deformation is very intense and the photograph shows two overturned folds. The deformed zone is about 20 cm. thick and is overlain and underlain by horizontal undeformed varved beds. It is thought that deformation of this origin may take place on a much larger scale in the sea due to shove of large icebergs.

OTHER DEFORMATION OF SEDIMENTS NOT DUE TO DIASTROPHISM

Deformation Due to Recrystallization

Recrystallization produces deformation if the new materials formed have greater volume than the old. It may be caused by change of aragonite to calcite, a change involving a volume increase of 8.35 per cent, or of anhydrite to gypsum with an increase in volume of nearly 63 per cent.

FIG. 73. Deformation referred to recrystallization. ½. Gypsum and anhydrite of Hillsboro, New Brunswick. (*Photograph by R. E. Rettger.*)

This type of deformation is beautifully shown in the Stassfurt deposits, where certain complex structures, have been termed *enterolithic* from resemblance to intestines. The structures produced through change of anhydrite to gypsum are excellently shown in the Hillsboro, New Brunswick, anhydrite and gypsum deposits, where the deformation is in the nature of miniature geosynclines and geanticlines of which some layers are independently folded and faulted with respect to others. The geanticlines and geosynclines have dimensions of 12 to 20 cm. (Fig. 73). The evaporite-bearing strata of the Permian of Kansas contain similar structures in abundance but also much larger anticlines and synclines, which are considered in the next paragraph. No deformations of any moment resulting from change of aragonite to calcite is known.

Deformation Due to Solution, Mining, and Quarrying

The removal of materials from mines and quarries is ultimately attended by some collapse and deformation of overlying and adjacent materials. Similar deformation attends the collapse of caves which are due to solution. Solution acting on such soluble substances as rock salt may remove very large quantities of material and be attended by considerable deformation of overlying strata. The evaporite-containing strata of some areas of central Kansas are greatly deformed, and anticlines and synclines of widths as great as 50 meters and heights of 6 or more meters ascribed to solution have been formed. Much minor deformation in the gypsum and adjacent strata is referred to change of anhydrite to gypsum.

CRITERIA OF CONTEMPORANEOUS DEFORMATION

1. Secondary cleavage bears no relation to contemporaneous deformation, as none is developed, and most parallel orientation of minerals is related to bedding.
2. Deformation is confined to zones between undisturbed zones.
3. Contemporaneous deformational structures may be beveled above or below, and the relations are those of local nonconformity.
4. Structures are simple to complex, and reverse structures are present in the same beds within short distances.
5. Cavities made by the deformation become filled with sediments of the disturbed zones and not by materials introduced in solution.
6. Faults have blurred contacts between the blocks disturbed and beds may show decided drag.
7. Rigid objects such as shells and clastic particles are not deformed.

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SPRING PITS

Springs are present on many beaches because of intersection of the water table with the surface. The water rises through the materials composing the beaches and forms low, but broad domes or mounds around places of egress (Quirke). These have gentle slopes, rise several centimeters above the general level, and the summit of each has a bowl-shaped depression that is 10 to 15 cm. in diameter. This passes downward into a tube. Domes approximate ½ to 1 meter in diameter. The coarsest materials are sands, which are found about the bases, and the finest sediments are at the top around the bowl-shaped depressions. Spring pits, as these structures are termed, form on beaches to a short distance above and also below the surface of the water. They should be preserved in the geologic column, but none has been recognized.

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MUD CRACKS

Mud cracks (sun cracks, shrinkage cracks) form as sediments lose contained water. The cracks bound polygons, which vary in number of sides and dimensions of angles between the sides. Directions and spacings of cracks vary with character and thickness of sediments, rate of drying, distribution of foreign matter in the muds, extent of stratification, salinity of waters of deposition, and character and water content of underlying sediments. Cracks are rarely straight, and polygons may be bounded by as few as three and as many as eight sides. Polygons with three to five sides are those most abundant. The often-repeated statement that mud-cracked polygons have six sides has little basis in observation.

Mud cracks form in all sufficiently cohesive sediments that are exposed to drying air, and the occurrence of fossil mud cracks is generally considered to indicate that such conditions obtained at the times of formation. It is possible that this does not always hold, as submerged muds may crack under some conditions.

Counterparts of mud cracks are formed on the bottoms of strata deposited immediately above mud-cracked surfaces. These may be in sediments in which mud cracks could not form, and they differ from the originals in being ridges instead of depressions.

The positions of mud cracks in sediments are determined by objects or conditions in the muds that decrease cohesion. Thus holes, thin places, sticks, hairs, and straw and other vegetable matter determine places of cracking. Holes are likely to determine several cracks that radiate outward. Cracks extend themselves along the places of least cohesion. They have many directions, are usually curved, and intersections produce polygons. Cracks first formed tend to be longest, deepest, and widest. Later cracks generally end where they intersect a crack of earlier formation. Homogeneous muds of uniform thickness and with very little or no foreign materials may crack so as to produce hexagonal polygons (Longwell). The formation of these has been explained as due to the development at numerous places of three cracks radiating from points and being separated by angles of 120 deg. Intersection of cracks thus oriented would lead to hexagonal polygons to a greater or less degree, but no one is known to have demonstrated that hexagonal polygons do form in this way (Fig. 5).

Width of cracks is determined by spacing, thickness and character of the muds, abundance of foreign materials in the muds, and order of development. The cracks formed first tend to be widest, wide cracks are the rule in thick-bedded muds, and closely spaced cracks in thin-bedded muds. Calcareous and sandy muds tend to have closely spaced cracks. Slow drying is said to produce close spacing, rapid drying wide spacing. Muds with an abundance of stems and other vegetable matter have closely spaced cracks (Fig. 74).

Mud-cracked polygons may turn up or down on the edges or do neither. What takes place depends upon the rate of drying or other extraction of water on different parts of a polygon. If extraction of water is greater on the upper side of a polygon the edges turn up, and if the extraction of water is uniform throughout, the surface remains flat. If muds are finer grained on top than below, the edges of polygons turn up; if finer grained below, the center of the polygon rises (Bradley). Muds deposited in fresh waters usually turn up on the edges, those of ordinary sea water tend to remain flat, and those deposited in highly saline waters tend to produce polygons that turn down on the edges. The dried muds of the latter also contain minute crystals of salt and pipelike passages that are not present in muds deposited in fresh water. Highly saline muds on drying also expel salts contained in the muds. These are deposited on the surface or in the

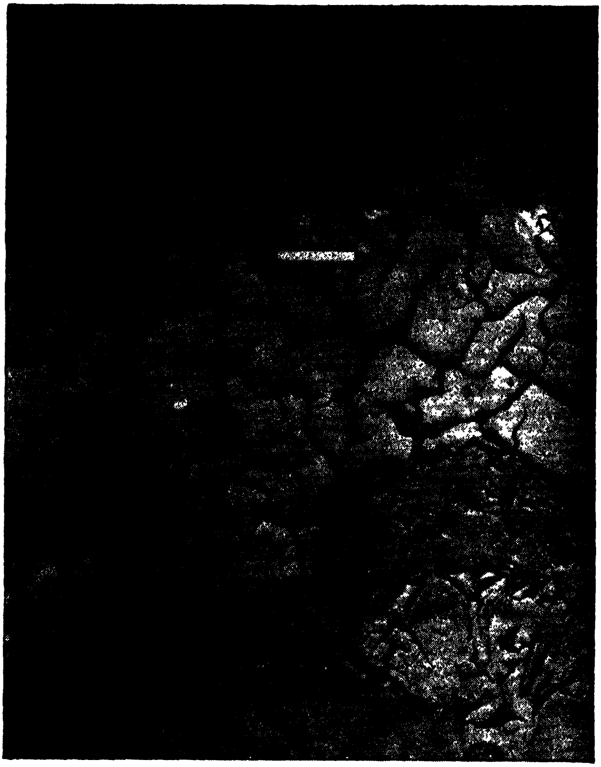

FIG. 74. Mud cracks developed in a tank. The muds are thickest at the top of the picture where the spacing is widest. Thin muds produce close spacing and in the lower right-hand corner, where the muds are thin, very fine-grained mud curls developed. It should be observed that all cracks are more or less curved.

surface materials, and this may lead to complete breaking up of the muds and their reduction to small particles and even to dust.

Wetting of mud-cracked polygons leads to slumping of marginal turned-up edges, and the middle parts may be left with circular outline. Freezing of slightly wet polygons and crystallization of salts in marginal parts may produce a similar effect. Some mud-cracked polygons when made wet do not slump on the margins. These contain substances that do not readily absorb water and soften and hence retain shapes for some time after submergence. Submerging waters may deposit sediment beneath the turned-up edges, and polygons in this position are thus preserved. This degree of preservation, however, seems best done if sediments are deposited from the atmosphere.

Cracks which are very like those made in drying muds are sometimes made when muds freeze. Cracks due to expansion instead of contraction may be produced beneath water if such materials as dry bentonite are buried beneath wet mud. This is not rare. A dry mass of bentonite on being wetted becomes impervious on the outside, thus protecting the interior which may be very dry. The interior later becomes wet and expands, with cracking of overlying muds. Cracks may also be produced in muds beneath water if the water in the muds drains out at the bottom.

Mud cracks are more abundant in continental than marine deposits. In marine sediments the average opportunity for drying is not good except in the upper part of the tidal zone and in marine playas. The tidal zone of most existing seas is narrow, and, correspondingly, chances for formation of mud cracks are small. At times of great development of epicontinental seas and very mature surfaces on the lands there may have been wide intertidal areas subject to mud cracking. During certain seasons of the year, vast areas like the Rann of Cutch of today may have been submerged for several months at a time and later exposed as a *marine playa* covered with mud cracks in infinite variety.

There is great development of mud cracks in some desert playas, but in others crystallization of salts breaks up the muds so as to destroy both cracks and polygons.

The finest development of mud cracks is found over floodplains and deltas of semiarid or treeless regions. Submergence of such floodplains and deltas during flood seasons deposits layers of mud. Retirement of waters leads to formation of a maze of crisscrossing cracks and polygons of great variety from widely spaced cracks and large polygons in thick layers of mud to closely spaced cracks and mud curls in thin-bedded muds.

Mud cracks have additional geological significance besides that of serving as indicators of environments. Repeated cracking of sediments may result in destruction of stratification and transformation of muddy sediments into a somewhat homogeneous nonstratified unit or into a desiccation breccia.

Mud cracks may also serve as keys to complex structures. Every mud-cracked layer has counterparts of the cracks in the form of ridges on the bottom of the immediately overlying layer, which merge without change of composition into the beds of which they are a part. The original cracks are filled with sediments which are like those of the counterparts and different from those of the bed. The ridges indicate the bottom of a bed, the original the top of one, and thus it is readily determinable whether or not strata have been overturned.

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CLAY GALLS

The mud curls or cylinders formed on cracking of thin layers of coherent muds are readily transported by rolling, and thus they may be deposited in aeolian or aqueous sediments at considerable distances from sources. They may flatten in aqueous sediments and become a bleb or lenticle of clay, known as a *clay gall* (Ger, *Tongallen*), in the midst of sediments of another character. Deposition with aeolian sediments may lead to filling of the curls with sands or silts and preservation of the filling as a stemlike structure surrounded by clay.

It has been pointed out that clay galls may form on some tidal flats from wet muds. A sand tidal flat at high tide may receive a deposit of a thin sheet of mud that between tides acquires considerable rigidity. The returning tide may tear the mud sheets into irregularly shaped pieces and carry these to various sites of deposition. Sands and muds over tidal flats may contain more or less organic binding matter, and the continuity of the surface layer is often broken by crawling organisms, so that thin sheets of mud are easily detached. Microorganisms feed on the organic matter in the muds and form gas that causes flakes of the mud to float when the tide returns. These flakes have areas ranging from 1 to 10 sq. cm., and, buoyed by the gases, they may be floated far from places of origin and deposited with other kinds of sediments.

Clay galls have been given the significance of dry to arid conditions at places of origin, but this can not be considered established until it has been demonstrated that they were originally mud curls. If preservation is in the form of cylinders, dry conditions also existed at the places of deposition.

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MUD PEBBLES AND BOULDERS

Mud pebbles and boulders usually form from wet mud dislodged from cliffs and stream channels. They are rarely, if ever, due to union of suspended particles of mud (Gardner). Some form from dry chunks, which after being wetted on the outside become impervious, and entrance of water into the interior is retarded. Ultimately water penetrates the interior, which expands and compels cracking of the outside.

Rolling leads to spheroidal, ellipsoidal, and cylindrical shapes. This may also produce increase in dimension through adherence of muds, sands, pebbles, shells, and pieces of organic matter. A concentric structure may thus be acquired. Boulders studded with small pebbles have been termed *pudding balls*, or *armored mud balls* (Bell).

Mud pebbles and boulders range in dimension to balls of 30 cm. in diameter and cylinders of equal length. Most are 1 to 10 cm. in diameter (Fig. 75).

Mud pebbles and boulders may become incorporated in sediments of any character in both marine and continental environments, as the low specific gravities permit them to roll in waters of low tractional competency. After burial beneath other sediments they tend to become flattened to disk and other shapes.

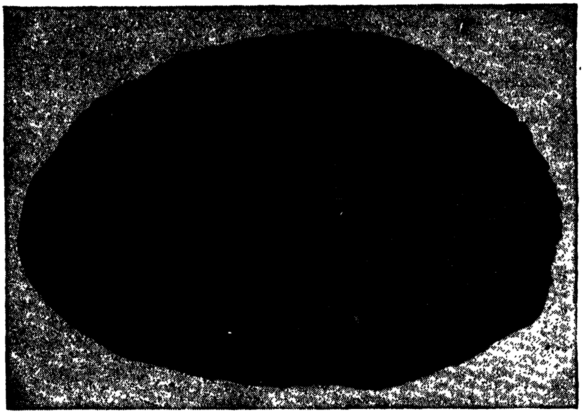

FIG. 75. Armored mud ball, collected on the shore of Lake Michigan. Consists of clay on the inside and is studded with pebbles on the outside. Dimensions are 16 × 8 × 4 cm.

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CONCRETIONS

INTRODUCTION

Concretions are aggregates of inorganic sedimentary materials in other sediments. Many have a nucleus, and internal structure is very commonly concentric. Oölites and pisolites are probably the most abundant concretionary structures, but these are reserved for separate consideration as they have different sedimentational significance.

Concretions are present in all unmetamorphosed sedimentary formations from the oldest to the youngest known. They originate in different ways and thus have various sedimentational significance.

Concretions have a wide range in composition and are rarely composed of a single substance. Calcite, silica, hematite, limonite, siderite, and pyrite (marcasite) form most concretions. Concretions are also composed of gypsum, barite, aragonite, witherite, manganese oxide, calcium phosphate, fluorite, phosphate, and bauxite.

Calcite concretions in the formational stage have been dredged from Auckland Harbor, New Zealand, and are now forming in the soils of many dry and semiarid regions. Calcite is the chief composing material of concretions present in shales. Concretions of aragonite are very rare.

Concretions of silica are mostly in the form of flint and chert. A few are composed of chalcedony and many consist of siliceous silts and sands cemented by silica or carbonate. Chert and flint concretions are commonly found in carbonate rocks. Those composed of siliceous silts and sands cemented by silica or carbonate are in silts and sandstones. Brownell has described concretions of quartz in deposits of gypsum and anhydrite.

Iron oxide concretions are very common in many sandstones and consist of quartz sands cemented by hematite or limonite. Iron oxide concretions form in some lakes and soils, particularly lateritic soils. The "shot soils" of warm latitudes are so named from the presence of small iron oxide concretions of the dimensions of shot. Iron oxide concretions, thought to be largely due to the decomposition of pyrite, marcasite, and siderite, are present in some limestones.

Siderite concretions are usually found in clays containing considerable organic matter. They are almost always impure, and the iron in the peripheral part may be in the form of oxide.

Pyrite and marcasite concretions are mainly aggregates of crystals. Structure is frequently radiate, and the concretions in these cases are spheres. They are common in sediments containing much organic matter and are very abundant in some paralic coals and marine black shales.

Gypsum concretions are occasionally common in shales and sandstones. Occurrences are in the form of porous balls of small dimensions, perhaps 5 cm. being a large maximum diameter.

Barite concretions form the "petrified roses" and "barite dollars." These are crystal aggregates of barite cementing the sediments in which they formed. Barite nodules have been dredged from the sea bottom, and barite oölites and pisolites have formed in oil wells of Texas. Witherite concretions are rare, as are also those composed of fluorite.

Manganese oxide in the form of wad or psilomelene forms concretionary bodies in secondary deposits of manganese, and nodules of manganese oxide are not uncommon on the present sea floor. Many of the latter consist of other sediments cemented by manganese oxide, and most contain nuclei, which may be pieces of rock or material of organic origin. Bauxite is commonly in the form of pisolites and larger concretionary bodies.

Dimensions of concretions range from a fraction of a millimeter in diameter, as in oölites, to large masses of spheroidal shape, with diameters as great as 4 meters, and loglike shapes more than 30 meters long. The largest concretions are thought to form in sandstone, where they largely consist of sands cemented by introduced cementing materials. Large concretions are also present in shales. Concretions in limestone, commonly composed of chert and flint, attain diameters of 2 or more meters.

The shapes of concretions vary within wide limits. Some resemble animals, plants, and even articles of human construction, as rings, hatchets, bombs, etc. Most have spheroidal, ellipsoidal, or disk shapes. However, cylindrical concretions are not rare and many concretions have shapes difficult to classify. Some spheroidal concretions are so extremely perfect as to seem to have been artificially constructed. Cylindrical concretions resemble fingers, roots (rhizoconcretions), logs, and even human beings.

Loglike bodies with internal structure arranged in cylindrical form and standing in vertical position are present in the Cambrian sandstones of Ontario. The "logs" are composed of quartz sands like those of the sandstones in which they are enclosed. They range in length to 20 ft. and have diameters which range to 10 ft. Bedding of the enclosing sandstones ends abruptly on the outer surfaces of the "logs." Hawley and Hart ascribe the origin of the "logs" to rising currents of water at the times of deposition of the sands. These currents are considered to have destroyed bedding at the places of rising. Later cementation made them.

The surfaces of concretions are equally as varied as shapes. Some have very smooth surfaces and are sharply defined from the enclosing rock. Others have rough and irregular surfaces, and many of these merge into the enclosing rock. Some concretions have slickensided areas on the upper side, which are due either to sediments slipping down over concretions or to concretions growing upward into sediments. The surfaces of many are covered with short to long protuberances and nodes, and some have symmetrically arranged structures that seem as if carved.

The internal structure of concretions may be concentrically laminated, radial, or amorphous. Some concretions have concentrically laminated internal structure with radiate structure in one or more laminae, and in others radiate structure extends from center to periphery and there may or may not be concentric structure. Pyrite and marcasite concretions are probably more often radiate than otherwise (crystal aggregates of pyrite and marcasite are not included). Calcite concretions are frequently radiate, and so also are gypsum and rarely limonite. Some concretions have horizontal laminations that usually, but in some cases do not, correspond with the laminations in the enclosing rock. These laminations are thought to represent planes of deposition. This form of lamination is very commonly shown in concretions in sandstones, but it also may be present in concretions in shales and limestones. Cone-in-cone structure is not uncommon in some concretions. Concretions without apparent internal structure may have the same color throughout, or there may be a mottled arrangement to the colors. Concretions with laminated structure usually have color distribution corresponding to the bands. Concretions composed of concentric laminae of different composition may have the most soluble of the composing substances removed by solution, leaving a central part detached from an outer portion, forming what are known as *rattle stones* or *Klapperstein.* Many concretions of iron oxide are filled with loose sand that rattles on shaking. The sands are usually little colored, and either they were never parts of the concretions or they have been decemented by solution.

Nuclei are present in many concretions, but most do not seem to have any. Nuclei consist of particles of different minerals, entire or broken shells, insects, crustaceans, fish, small reptiles or amphibians, excremental particles, and leaves, fruits, and other vegetable matter.

OCCURRENCE OF CONCRETIONS

Concretions seem to be present in all kinds of sedimentary rocks except most evaporites and the bedded cherts and flints. Concretions in mudstones in approximate order of abundance are composed of calcite, calcite and clay, pyrite or marcasite, siderite, calcium phosphate, and gypsum. Black marine shales and paralic coals are characterized by pyrite and marcasite concretions. Gray shales containing considerable organic matter contain clay ironstone (sideritic) concretions. Sandstone concretions largely consist of sands cemented by (in order of abundance) iron oxide or hydroxide, calcite and quartz, and less commonly by barite and manganese oxide. The concretions in limestone and other calcareous rocks are mostly chert and flint.

Concretions commonly lie along bedding planes and more rarely within beds. Those on bedding planes are in shallow depressions in the underlying beds. There may also be depressions in the overlying beds. Stratification planes of the enclosing rock may pass over concretions, terminate against them either abruptly or with a slight upward curve, or continue through them. Some concretions cross bedding planes. Concretions within beds usually lie along definite horizons, but they may be distributed irregularly. In general, the longest axis of a concretion is parallel to bedding planes, but exceptions are many. Most concretions in shales are along bedding planes. Chert and flint concretions in limestones are mostly along bedding planes, but they are also very common within beds. The concretions in sandstones seem mostly to have formed long after the formation of the enclosing rock, and, as sandstones are generally pervious, it results that many concretions in sandstones are within beds.

ORIGIN AND CLASSIFICATION OF CONCRETIONS

Concretions seem to be best classified on the basis of time relations to the enclosing rock and thus may be placed in two divisions, which are:

- Syngenetic or contemporaneous (primary), formed at the same time as the enclosing rock.
- Epigenetic or subsequent (secondary), formed after deposition of the enclosing rock.

A third group, termed *penecontemporaneous,* has been proposed to include concretions formed close to the surface of recently deposited sediments. These in most instances are best placed in the syngenetic division, as materials in solution or colloidal suspension probably play a considerable part in their formation.

Concretions at one time or another have been postulated to have developed through organic, physical, and chemical processes. The so-called concretions of organic origin include algal structures variously known as *Cryptozoa, Girvanella,* marlite balls, marl and lake biscuits, and other names. Also included at times have been such coralline structures as *Clathrodictyon.* The view is held that none of these should be considered concretions in spite of resemblances in appearance and structure.

The so-called physical concretions are the mud pebbles and boulders formed by accretion. These should not be considered concretions.

If the organic structures and the mud balls enlarging by accretion are excluded, the term concretion becomes limited to structures of chemical origin. It is not intended to be implied that organic processes may not be concerned in precipitation of the substances that form concretions, but that the concretions do not bear structural evidence that participation of organisms was responsible for the structure. Naturally, there are structures that become difficult to place, but that problem is inherent in all classification.

By definition, syngenetic concretions are formed contemporaneously with the sediments that enclose them, and the composing materials are precipitated from solution or colloids in suspension. Syngenetic concretions of calcareous materials are known to be forming in Auckland Harbor of New Zealand, concretions of manganese oxide in Ship Harbor Lake of Nova Scotia and on the present ocean bottom, concretions of iron oxide in some lakes and soils, and concretions of calcium phosphate and manganese oxide on the present sea floor.

Concretions are considered to be of syngenetic origin if the nuclei consist of well-preserved fossils, as those in the coal balls (calcareous concretions in coal) of the Lower "Coal Measures" of England and the concretions in the Pennsylvanian strata of the famous Mazon Creek locality of Illinois, and if stratification planes of the enclosing strata end abruptly or bend upward against them, particularly if laminae thicken adjacent to the concretions. If laminae curve over and also under a concretion one cannot be certain of syngenetic origin, as this may take place in concretions of epigenetic origin that develop between beds and in syngenetic concretions because of settling of beds beneath and around them. Slickensided surfaces develop in either case. The physical character of the enclosing rock also bears on the problem. If this is impervious, it may be considered that enclosed concretions are syngenetic. Size of concretions also has relations to origin. Very large concretions that were not formed as a consequence of replacement or cementation of the enclosing rock could hardly be formed by displacement without development of obvious structures connected with the displacement. If these are absent and the concretions are not due to replacement or cementation it may be concluded that they are of syngenetic origin. Concretions are very common along bedding planes, but they are extremely rare along joint planes, where they should be as abundant as along bedding planes if those along bedding planes are of epigenetic origin. It seems far more likely that such concretions are of syngenetic origin and were deposited in the interval between deposition of the underlying and the overlying bed.

Epigenetic concretions by definition form in sediments after deposition. Formation may take place in the sediments through redistribution of the composing materials while they are still soft, but sufficiently buried so as to be little influenced by overlying waters; at the time of consolidation; or after the strata have been uplifted and the more or less indurated sediments have entered the influence of ground-water circulation. It seems likely that concretions may be formed at any time in this sequence of events, but it is considered that most epigenetic concretions form either before the enclosing strata are consolidated or after they enter the influence of active ground-water circulation. It is known that many calcium carbonate and iron oxide concretions form in rocks and soils above or adjacent to the water table, and it is considered probable that many chert and flint concretions form as rocks are decomposed in weathering.

The materials to form epigenetic concretions must be obtained from the enclosing formation or from outside sources. Rates of growth are functions of availability of materials, rate of movement and solvent ability of the transporting medium, rate of precipitation, and porosity and permeability of the enclosing materials. Materials of low porosity and permeability, particularly the latter, are not promising hosts for formation of epigenetic concretions. This restriction is to a considerable extent removed after rocks enter the zone of weathering.

It is thought that sands and sandstones offer the best conditions for the formation of epigenetic concretions, and that conditions are less favorable in limestones and least favorable in mudstones.

Epigenetic concretions may be formed by cementation, replacement, or displacement. Formation by cementation requires sedimentary materials sufficiently pervious to permit migration of solutions. This limits formation of epigenetic concretions of this kind very largely to sands and silts and sandstones and siltstones. Waters circulating through pervious materials gather materials in solution and under favorable conditions deposit these around centers, cementing the materials there present. The iron oxide concretions so common in sands and sandstones were formed in this way, the iron probably being gathered from the enclosing sediments and transported either as colloidal iron oxide or as iron carbonate. The famous concretions of the Connecticut Valley clays were formed through local cementation of more porous and pervious layers by calcium carbonate deposited from solution. The thin disklike shapes are probably due to thinness of the layers of fine-grained sandstone to which the concretions are largely confined. The enclosing clays, silts, and sands are thought to have contributed the carbonate to the circulating waters (Tarr).

A variety of concretion formed by replacement originates through decomposition of some substance within a rock, as a particle of pyrite, marcasite, or organic matter. The products of decomposition travel outward from the center of decomposition and discolor the rock in banded fashion and perhaps produce some differential cementation in the bands.

Formation of concretions by displacement assumes that the force of crystallization lifts or compresses the enclosing, particularly the overlying, sediments. This is probably an easy task for small concretions but becomes extremely difficult in the cases of large ones, and if concretions have diameters of 0.5 meter or more, it means that at least 0.125 cubic meter must be displaced. This would unquestionably have effects on the enclosing rocks in the form of warping of strata both above and below, and as stretching is involved in the warping it can only be done by thinning or breaking the strata. The displacement necessitated by formation of small concretions can perhaps in most cases be cared for by compaction and recrystallization of, and subtraction of materials from, adjacent sediments, but large concretions necessitate lifting overlying beds, which in many cases can hardly be done by force of crystallization except in the early stages of epigenetic history when the sediments are on the sea bottoms and have a small overburden, or in the last stages when the enclosing rocks are in the zone of weathering. The common presence of cone-in-cone in concretions and slickensides on the upper surface have been interpreted as results of pressure produced by replacement, but most of the cone-in-cone is on concretions that are best interpreted as of syngenetic origin, and the slickensides are equally well explained as due to sediments settling over the concretions.

Many concretions have been postulated to have been formed by displacement, as the large calcareous concretions in the Devonian at Kettle Point, Ontario, but the same concretions have been referred by others to syngenetic origin. The chert and flint concretions so common in limestones have been referred by some to displacement, and some of them seem certainly of this origin, formed in the early or late stages of epigenetic history, but many cannot be so referred.

Concretions formed by replacement involve removal of original materials and replacement by others precipitated from solution or coagulation of colloids. Removal is possible in soluble rocks but hardly possible in sandstones and mudstones. Some chert and flint concretions have been postulated to have been formed in this way, and it is thought to be known that this is the correct explanation for some of them. Some iron oxide concretions in limestones may also have been formed by replacement. Original structures of the rocks replaced should be present to a greater or less degree.

Epigenetic origin of concretions is indicated by the presence of structures present in the enclosing rocks, as these would be preserved to some degree by cementation or replacement. Distortion due to growth of the concretion by displacement should be indicated in the enclosing sediments. The physical character of the enclosing rock should be a factor of importance in determining the method of origin.

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SEPTARIA

Many concretions composed of calcite and rarely of other substances have cracks (veins) extending inward from the periphery that are partially or completely filled with many varieties of minerals, among which are calcite, barite, selenite, marcasite, pyrite, arsenopyrite, millerite, galena, sphalerite, chalcopyrite, and others. The minerals may be in the form of crystals if the veins are not filled. The cracks (veins) narrow toward the center, and maximum widths on the periphery are usually less than 5 cm. Concretions of the structure described are known as *septaria.* Many septaria have slickensided surfaces, and some have cone-in-cone structure. Septaria not uncommonly resemble fossil turtles, particularly if the cracks or veins are symmetrically arranged.

Solutions acting on septaria may remove the fillings of the cracks, in which event the septaria fall apart, or may remove the material between the cracks, thus leaving a skeletal structure made up of the veins. Similar skeletal structures of other veins have been termed *melikoria,* and Burt has suggested that this term might also be applied to the vein skeletons of septaria.

The origin of septaria has occasioned considerable differences of opinion. Richardson's views seem to be in accord with the facts. He postulates that a septarian at first was a concretion with a colloidal central interior. This ultimately crystallized, and water was expelled. Crystallization and expulsion of water resulted in shrinking of the central interior and led to its cracking. Minerals were ultimately deposited in the cracks. These forced the walls apart and extended cracks to the periphery. Other views are those of Todd who suggested that the exterior expanded through internal growth and pulled the interior apart, and Davies that the interior expanded to produce the cracks. Richardson's view involves expansion of the interior through growth of minerals in cracks but by an entirely different mechanism from that of Davies.

As septaria are almost entirely limited to shales, in which in some areas they have wide distribution, the postulate of a colloidal center is easily made.

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GEODES

Geodes are hollow nodular bodies enclosed in sedimentary rocks. Exteriorly, they bear considerable resemblance to concretions. Surfaces are usually not smooth, and the internal cavity is lined with crystals or crystalline materials. Geodes may form in any sedimentary rock through deposition of material from solution in circulating waters upon the walls of somewhat spherical or ellipsoidal cavities. A common method of origin of the cavities in some formations is by deposition of material from solution, chiefly silicon dioxide and calcium carbonate, along structural and fracture surfaces of shells (Bassler). This deposition leads to enlargement of the cavities and ultimately may line them with minerals of which many are in crystal form. The geodes of this origin are commonly enclosed in calcareous shales or highly argillaceous limestones, and every stage in the formation of geodes from crinoids, gastropods, brachiopods, and shells of other organisms has been observed. Common minerals in geodes are quartz, chalcedony, and calcite, but dolomite, barite, celestite, sphalerite, galena, sulphur, and many other minerals have been found. If the materials deposited on the walls of cavities are less soluble than the enclosing rock, the latter may be removed and leave the filling with its internal cavity free. This takes place very commonly in argillaceous limestones, and the geodes in shales free easily when the shales soften.

Reference

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CONE-IN-CONE STRUCTURE

DESCRIPTION

Cone-in-cone structure is generally formed in argillaceous calcitic limestones and consists of parallel nests of cones fitting into each other, and there are occasional single cones. Cones range in height from very small to 20 cm.; most range from 10 mm. to 10 cm. Basal diameters depend upon heights and dimensions of apical angles of cones. Heights and basal diameters in some cases are nearly equal; most have heights much greater than basal diameters. Apical angles range from 15 to about 100 deg.;

most are between 30 and 60 deg. Sides of cones are ordinarily straight, but diameters of some cones increase so very rapidly toward the bases that sides are concave outward. Fluting or ribbing generally covers sides of cones, and many have the fine striations and polished surfaces characteristic of slickensides. All markings on the surfaces of cones have counterparts on the interior surfaces of enclosing cones of which the insides are termed cone cups. Conic scales are laminae that extend over only parts of cones. The sides of cone cups have more or less discontinuous annular ribs which are parallel to the bases and are usually less than a millimeter wide but may be 5 to 6 mm. wide. These have smaller counterparts on the surfaces of cones. Dimensions are greatest adjacent to bases of cones and near tops of the cone cups, and they become smaller with approach to the apexes of cones and bottoms of cups. The width and abundance of annular depressions and ridges seem directly related to the clay content of cones. The depressions in the cone cups are filled with clay, and thin films of clay are common between cones and between cones and conic scales (Fig. 76).

Most cones are composed of fibers of calcite. The fibers are rudely circular in cross section and are not known to be terminated by crystal faces. They are commonly parallel to the cone axis, but some may be inclined, and there may be a tendency on the surface of cones for the fibers to become parallel to the cone surface, particularly the ends of fibers. Fibers on the apexes of some cones are more or less bent. Parting planes are present in some cone-in-cone layers, and both fibers and cones end at these planes.

Cone-in-cone in manganiferous siderite (Hendricks) occurs in Arkansas, and gypsum cone-in-cone is present in Comanchean strata of south central Kansas. The latter seems to have been formed by alteration of calcium carbonate. Apparent cone-in-cone structure is present in coal of Monmouthshire, England, and southern Wales. The cones are united so that the bases of the cone laminae form curved and somewhat concentric parallel ridges. The cones average 30 to 40 mm. in height and have apical angles of 36 deg. The sides are slickensided and have conic scales. Two sets of cones are present with apexes directed in opposite directions, and the adjacent bases are separated by a thin layer of coal or clay. Bartrum (1941) reports "cone-in-cone" in some New Zealand coals and Gage and Bartrum describe another occurrence in New Zealand. It is questionable if these occurrences in coals are real cone-in-cone. Percussion cone-in-cone has been found in quartzite, but this does not seem to be in the same class as cone-in-cone composed of calcite, although some factors concerned in its origin may be the same.

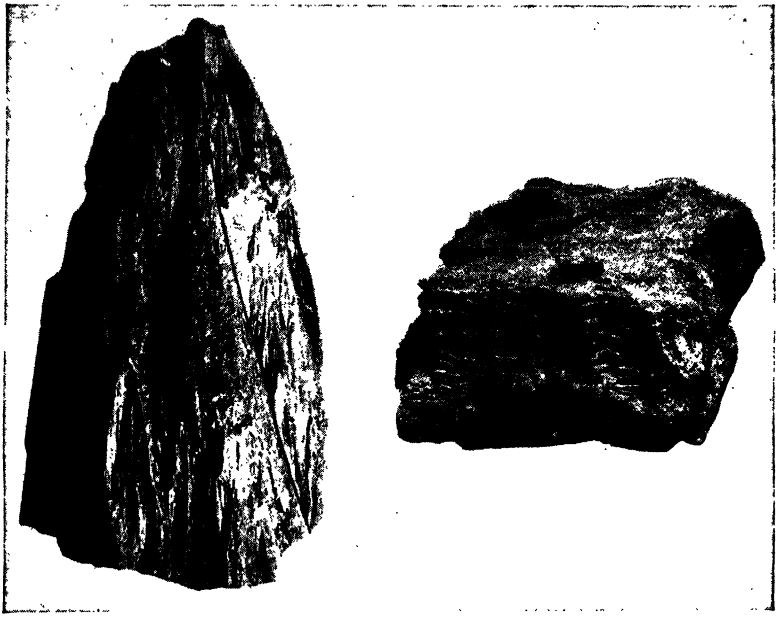

FIG. 76. Cone-in-cone structure. The cones in the larger specimen are about 12.5 cm. long. The smaller specimen shows one cone seemingly in process of being forced outward. Specimen from the Lower Cretaceous at Langley, Ellsworth County, Kansas. The annular ridges and depressions are shown in the smaller specimen.

Cone-in-cone layers are usually associated with mudstones, and there is also a common association with concretions, although most concretions are without the structure. A cone-in-cone layer may be on any part of a concretion. Most cone-in-cone layers have a somewhat lenticular character, but there is a layer in the Lower Cretaceous of Kansas that extends over many square miles. This, however, thickens and thins in lenticular fashion. Layers may be single, in which case the apexes are directed downward and the bases are above, or double, with the apexes of each part directed toward each other. Cones on concretions may have the apexes directed in any direction.

Cone-in-cone structure is found in the rocks of every geologic system from the Middle Cambrian to the Tertiary. Distribution is world-wide.

ORIGIN

The origin of cone-in-cone structure has not been definitely determined. A few have advocated that the structure developed as a consequence of gas rising through sediments. This view probably has little present support. Many have postulated that the structure is connected with crystallization, particularly as calcite at times crystallizes in fibrous form. This obviously could not apply to coal. One phase of this theory explains the cones as resulting from radial crystallization of calcite around an axis, the clay between the cones being expelled as the calcite crystallized. The latest advocate of the crystallization theory is Richardson, who postulated that cone-in-cone structure arises as a consequence of internal stresses that are produced as calcite crystallizes in fibrous form under the load of the overlying sediments. The overlying load creates a vertical stress, and lateral stresses are assumed to develop as resistance to lateral growth of the calcite fibers. As consequences of these two stresses conical surfaces of maximum shear are produced, and the apical angles of the cones are determined by relations between the two stresses. The inclination of the rhombohedral cleavage to the calcite fibers makes the calcite extremely sensitive to the conical surfaces of shear. After the shear planes are established there is little chance for formation of fibers crossing them.

The radial-crystallization form of this theory may be disregarded, as the fibers are not in radial arrangement but are mostly parallel to the axes of the cones. Fibers in a few cases are parallel to shear planes, but this seems to be a later adjustment. Richardson's view of lateral stresses occasioned by growth of fibers does not seem possible, as Tarr has shown that cones are later than the fibrous layers in which they occur, and hence growth of the fibers could not have produced lateral stresses that would have aided in formation of the cones. Shaub advanced the theory that cone-in-cone is due to contraction, settling, or volume shrinkage during the dewatering of highly saturated, porous, fine-grained sediments deposited in a loosely packed condition. He noted funnel-shaped depressions in river muds which he thought might be responsible for cones. The present author sees little support for the Shaub theory.

The pressure-and-solution theory of origin of cone-in-cone is the most acceptable at the present time. Most cone-in-cone structure shows movement of cones with respect to each other, in the presence of slickensided surfaces between cones and cone cups and between cones and conic scales, by penetration of an inner cone into an outer without breaking the surface of the latter, and by a depression on the base showing the position of the inner cone. Solution is indicated by the presence of clay between cones and between cones and conic scales. The fibrous character of the calcite may have aided in formation, as would likewise the rhombohedral cleavage of calcite.

The source of the pressure is assumed to be the load of the overlying sediments, but it has been suggested that it may be found in the change of aragonite to calcite, although no aragonite seems ever to have been found in cone-in-cone; the growth of concretionary structure with which much cone-in-cone is associated, although most cone-in-cone has no relation to concretions; or diastrophism, in spite of the fact that most cone-in-cone is in strata that have not been disturbed. Moreover, only displacement concretions develop pressures, and the known concretions bearing cone-in-cone seem to be of syngenetic origin. Crystallization of calcite has also been suggested, but, as already stated, the cones formed after crystallization, as is shown by the fact that cone surfaces cut across the fibers, thus proving that the fibers preceded. The load of overlying sediments remains the only source of pressure.

The various features of cone-in-cone structures accord with the idea that differential vertical pressure due to the overlying sediments is perhaps the major factor in production and that the cleavage planes of the calcite composing the fibers and solution are secondary factors. The features considered pertinent to this explanation are (*a*) cone-in-cone is largely in horizontal or little-inclined strata; (*b*) cones are mostly on the upper sides of cone-in-cone layers and have the bases upward, and if cones are on the undersides they are smaller and have the bases downward; (*c*) apical angles of the most perfect cones approximate those formed in testing the crushing strength of materials (70 to 110 deg.) and accord well with the rhombohedral cleavage of calcite (106 deg.), and sharp blows with the hammer on fibrous calcite produce percussion cones that are similar to those in cone-in-cone; (*d*) solution and rearrangement of fibers are present along minor faults that cut the fibers; (*e*) slickensided phenomena are present between the cones and between cones and conic scales; and (*f*) solution is shown in the clay in the annular depressions and in the films between cones and between cones and conic scales.

The load increased as the sediments accumulated. It was constant, continuous in application, and of great magnitude, 1,100 lb. per square inch for each 1,000 ft. of average rock. It is not to be expected that the pressure would be uniform over the surface of a fibrous layer, as that would be precluded by variations in character of the immediately overlying materials, the underlying materials, the surface and thickness of the fibrous layer of calcite, and the rhombohedral cleavage of the calcite fibers. The character of the cone-in-cone would also be influenced by variations in the fibers, as some are thick and some thin, some prismatic, others pyramidal, some extending from one side of a layer to the other, and others extending only part way across. The pressure of the overlying rock developed conical shear planes in the fibrous calcite, the apical angles of the cones being somewhat controlled by the size of the calcite fibers, the vertical joint planes between the fibers, and the rhombohedral cleavage of the calcite.

Solution began after establishment of shear planes, calcite was removed, and an insoluble residue of clay was left. This aided movement along the shear planes, as the clay served as a lubricant. Where fibers were cut by shear planes, solution was most active, because the fractured ends were under pressure, particularly the upper ends that appeared on the inside of the cone cups. This solution formed the annular depressions in the cone cups. Slickensided features developed as the cones slid over each other, and conic scales represent either imperfect cones or remnants of cones left after solution. Removal of material between fibers permitted them to rotate to parallelism to shear planes, particularly at the ends. This would be most apparent on the apexes of cones. The bluntly rounded apexes of some cones are readily explained as results of solution. Deposition of calcite may also take place along shear planes, particularly after pressure is released through erosion of overlying sediments. This may result in driving cones in the direction opposite to that of formation and in forcing the *base of a cone to protrude beyond the surface of a cone layer* (Fig. 76). Actual deposition of calcite along the shear planes to a thickness of more than 1 mm. is known.

The preceding statement of origin should not be considered as completely demonstrated. The theory, however, seems to accord with and explain the facts. Modification may be required as new facts are discovered.

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STYLOLITES

DESCRIPTION

Stylolites (also termed epsomites, crystallites, and crowfeet) are vertically striated columns, pyramids, and cones of various heights, widths, and shapes on bedding planes and less commonly on planes of other origin. The stylolites on the upper surface of any bed fit into those on the under surface of the overlying bed so that each is a counterpart of the other. The two constitute a stylolite seam. Separation of a stylolite seam shows an extremely irregular surface that resembles in miniature a mountainous region except that little system obtains in arrangement and orientation of the elevations and depressions and none in the heights, depths, and shapes. A stylolite seam may pass laterally into a bedding plane on which no stylolites exist.

Seams may extend from a few millimeters to many feet. Several seams may unite, which in some cases involves cutting across laminae, which then end without deflection against the stylolites. Examples of younger seams cutting across older stylolite seams are known. Most seams are horizontal or nearly so, but seams at angles of 45 deg. or more are known. Formation seems to be response to vertically acting factors.

Each stylolite seam is characterized by the presence of a thin film of clay that is thickest on the summits of the columns and in depressions. This clay has the composition of the insoluble residues of the rocks enclosing the stylolite seam, and its thickness seems related to quantity of insoluble residue in the rock penetrated by the elevations and to the extent of penetration. Thickness of a cap may range to 10 mm. or more.

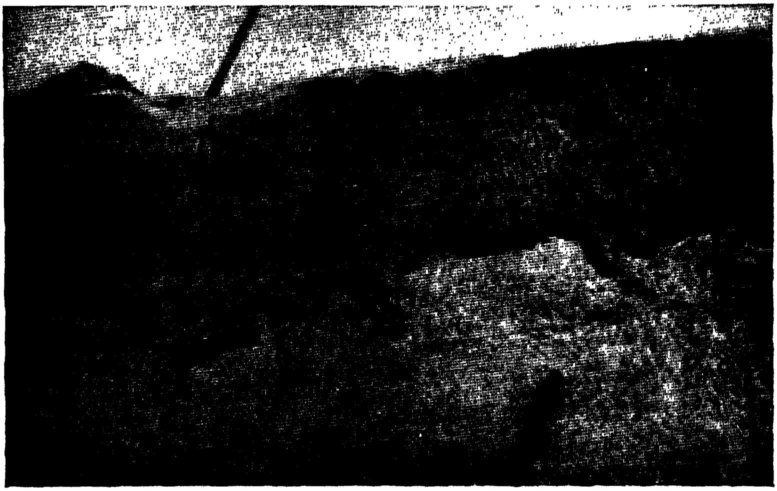

FIG. 77. A stylolite seam from the Salem limestone of Lawrence County, Indiana. The clay parting is plainly visible, and the upper and lower strata are distinctly lithologically different. The longest column is about 22.5 cm. (*After Stockdale.*)

Stylolites are largely confined to limestones, both calcitic and dolomitic, but rare occurrences of what seem to be stylolites have been seen in sandstone, quartzite, gypsum, shale, and chert (Sloss and Feray). Very impure limestones have few stylolites, but slightly impure ones may have them in abundance. Thus the fine-grained Niagaran dolomitic limestones around the Michigan Basin and the Mississippian limestones of the Mississippi

Valley have them in abundance, but they are extremely uncommon in the impure limestones of the Anticosti and the Cincinnati regions (Fig. 77).

Heights of stylolites range from less than 1 mm. to over 30 cm., and most range from less than 1 to about 60 mm. and are commonly 3 to 20 mm. Very small stylolites have been designated suture joints (German, *Drucksuturen*). Tops of stylolites may be broad and flat, narrowly to broadly rounded, or single to many pointed. Shape seems to bear little relation to height. Most are straight, a few are curved, and examples of one stylolite cutting across another have been seen. Small stylolite seams penetrated by large stylolites of an adjacent seam are not uncommon.

The literature contains statements that a fossil shell caps each stylolite column, and it has been postulated that these shells are connected with the origin of the columns. Studies of thousands of stylolites by Stockdale, this author, and others have failed to find any basis for such statements. Perhaps in one of a thousand cases a shell may cap a column, and evidently the relation is purely incidental. Shells partially penetrated by stylolites are not uncommon.

ORIGIN OF STYLOLITES

Stylolites in early explanations of origin have been referred to the work of organisms, shrinkage cracking, erosion of soft sediments by rain before burial with no erosion at those places where shells or pebbles provided protection, and rising of gas through soft sediments. All these explanations may be dismissed as of no application. Careful studies seem to have shown that stylolites formed after consolidation of the containing rocks and that the responsible processes are pressure and solution. Solution is held to be directly responsible, and pressure directs and localizes solution with such modification as may be imposed by the character of the rocks involved. The pressure is derived from the load of overlying sediments. This has differential application along parting planes in the sediments, which are commonly bedding planes. The stylolites are developed along the parting planes through solution controlled by differences in the application of pressure and differences in the solubility of the rock. Most soluble parts are removed. This shifts application of the pressure and, correspondingly, the solution. The stylolites are thus mutually extended into the adjacent beds by removal of material ahead of them, and the insoluble residues of the rock dissolved remain as films over the seam surface and as caps on the summits of the stylolites. If the pressure is not of differential application and if all parts of the rocks involved are of equal or nearly equal solubility, solution would take place uniformly and stylolites would not be likely to form, but a plane to undulating seam marked by a film of clay representing the rock dissolved would develop. Stylolites in shale hardly fit into this explanation, but those in quartzite and sandstone may do so. The stylolites reported in shale, sandstone, and quartzite should have further study. There may be other factors concerned in stylolite formation that are not yet discovered.

The application of the pressure is largely vertical, and hence the stylolites have nearly vertical orientation, and if they have general inclination to the horizontal this is probably due to tilting after formation of the stylolites. Occasional inclination from the horizontal may be expected in any seam.

Shaub has advanced the theory that stylolites originate in unconsolidated soft sediments as results of slow differential plastic flow during compaction under load, that no solution is involved, and that the clay partings along stylolite seams represent original clayey beds deformed by the differential plastic flow. The present author does not consider the theory in accord with the features which are present in stylolites. He considers the Stockdale explanation the valid one.

That solution is responsible for stylolites seems proved by the composition of the clay seams being that of the insoluble residues of the adjacent rocks, by the thickness of the clay cap being in direct relation to the height of the column on which it rests, by penetration of fossils by stylolites, by penetration of older stylolite seams by younger ones, and by the undeflected ends of lamination and other structures at contact with stylolite columns.

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CURRENT MARKS

Current marks are irregular structures produced by erosion. They are excellently shown on uncovered tidal flats where the falling tidal waters erode numerous channels and leave uneroded areas as flat "plateaus" between channels. Channel fillings made on burial of such surfaces may resemble casts of logs. Current marks are also made in the beach zone on the lee side of obstructions. A small depression begins at the obstruction and extends downward toward the water. They are common on beaches in the tidal zone on the downshore side of pebbles, shells, etc.

MISCELLANEOUS MARKINGS DUE TO WAVES

Swash Mark

Waves breaking on a beach produce the swash, a thin sheet of water that glides up the beach until the beach-directed energy is lost. The place of highest advance is marked by a ridge produced by dropping of the load of the swash as the water comes to rest and some removal of material on the water side as the water begins to retire. This ridge, known as the swash mark, is composed of sands, mica flakes, fragments of shells, and pieces of seaweed and other vegetable debris. Heights of ridges approximate 1 to 2 mm. As waves differ in intensities, places of farthest advance of the swash are not everywhere the same. Swash marks are likely to be destroyed by succeeding waves, but with declines in wave violence or fall due to receding tide, waves rise less and less high on a beach, and numerous swash marks are left. These form an irregular pattern of anastomosing ridges with broad convexities toward the land and concavities and angles toward the water (Fig. 78). Swash marks have little chance of preservation, but they are said to have been seen in the Medina and Portage formations of New York. Their presence indicates a beach.

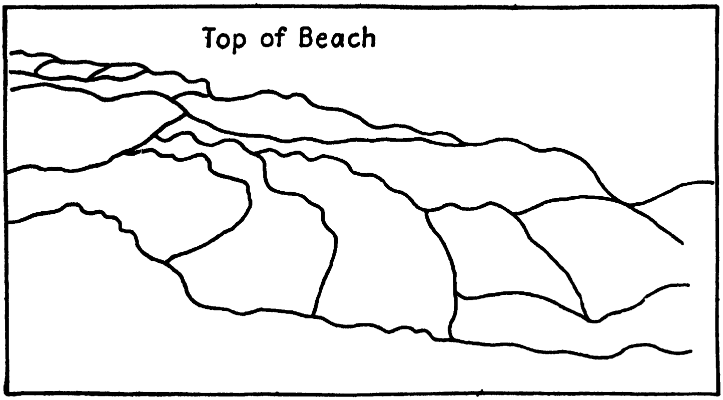

FIG. 78. Tracing of swash marks taken from a photograph of swash marks. Beach of Mingan Islands, Quebec, Canada. The water is at the bottom and the high part of the beach above.

RILL MARKS

Another common mark on beaches is that made by small rills that develop as high waves fall or tides recede. Muds and sands of beaches covered with water become saturated, and, as water levels fall, the water in the sediments drains to the surface at some lower level and becomes aggregated into small rills of dendritic pattern. Rills may also form in the tidal zone on the slopes of ripple marks after these become uncovered. Rills range from about 2 to 10 mm. wide and may be 0.5 meter or more long. They erode channels about 1 mm. deep and the pattern resembles plant impressions. The dendritic pattern may be about one-third as wide as it is long. Marks of this origin are thought to have been preserved in the geologic column, although chances of preservation are not good.

SAND HOLES

As waves retire from a beach they leave its materials saturated. Some of the water flows out between waves and forms the small rills described above. Air then takes the place of the water. Returning waves expel the air, which escapes in the form of small bubblelike masses, and the material again becomes saturated. Expulsion tends to be violent and to form small pits that are 7 to 8 mm. in maximum depth and a little less wide than deep. The pits have small raised margins, resemble rain-drop impressions, and are known as *sand holes*. They were first described as having been excavated by sand hoppers and it has been suggested that some tubes ascribed to *Scolithus* may have been formed in this way. Sand holes are extremely abundant on beaches composed of suitable materials, and falling waters leave them behind by thousands. They have little possibility of preservation, and it is not known that any are present in the geologic column.

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BEACH CUSPS

Beach cusps are large ripplelike structures that project outward from a shore and are composed of coarse materials in the range from sand to cobbles. The shape in plan is more or less that of an isosceles triangle of which the base is on the beach and the apex extends outward into the water and may be submerged. The angle at the apex is usually small, but it may be fairly large. It is nearly always acute. Sides are more or less curved, generally concave outward. Heights may be 1 meter or more. The highest point may be at any place on a cusp, so that the apex of a cusp may be lower than the highest point and a depression may be at the base. Lengths range from a few meters to 30 meters or more. Distances between cusps vary from a few to many meters (Evans). In some cases they are uniformly and in others irregularly spaced. The tendency is for the materials of a cusp to merge into those of the beach, but cusps may be composed of coarser materials than the beaches which they join. These structures form only on the beaches of large lakes and the sea, and the absence on small lakes indicates that waves are essential for formation. They obviously are consequences of wave erosion and deposition. Kuenen postulates that the swash erodes depressions and is deflected laterally to form the cusps, and the backwash forms submerged deltas outward in the bays between the cusps.

It is not certain that beach cusps have been recognized in the geologic column, but sharp-crested ridges with symmetrical profile in the Medina sandstone of New York may be of this origin. As these are shore features, the likelihood is that they would be destroyed by wave erosion as a sea advanced and by subaerial erosion as a sea retreated.

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SAND DOMES

Sand domes form on beaches composed of fine sands or muds of such character that the surface materials are poorly permeable and contained water does not readily escape. Air or steam collecting a few centimeters below the surface lifts the materials to form a small dome or blister. Sand domes may be seen on some beaches when the temperature is high, and they are readily destroyed by any perforation that permits the gas to escape. Sand domes can not be preserved in the geologic column.

Reference

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STRUCTURES MADE BY FLOATING OBJECTS

Floating objects of any kind that touch bottom produce grooves and furrows on suitable sediments. These markings have directions and dimensions proportional to the dimensions of the floating objects and the extent to which these touch bottom. Floating shells may produce small traillike grooves on soft muds of very shallow bottoms. Many of these strongly resemble trails made by worms and wormlike animals. Objects falling and rising with waves may contact the bottom only when in the troughs of waves, thus producing a series of pits or discontinuous furrows. Any object floating from shallow to deeper waters may form a trail only over the shallow bottoms. Under ordinary conditions the marks made by floating objects are on muds, but large floating objects may produce markings on sand bottoms. Most markings of this origin, except those made by floating ice, are made on bottoms of limited depths of which the maximum is perhaps about 3 meters.

Markings on muds would not be likely to be seen on exposure because of disintegration, but if burying sediments consist of sands, counterparts of the markings on the muds would be shown on the underside of sandstones. These have been seen on the undersides of Pennsylvanian sandstones of Oklahoma (Powers).

Overturned position of strata may be determined from these markings or their counterparts.

Reference

Powers, S., Strand markings in the Pennsylvanian sandstones of Osage County, Oklahoma, *J. Geol.*, 29 (1921), 66–80.

MARKINGS MADE BY BASALLY ATTACHED PLANTS

Plants growing on bottoms of water bodies or upon dry sand areas are swung around by movements of water or air, and, as branches contact muds or sands, markings are made that are arcs of circles. These may be seen in abundance over sand areas on land and over mud and sand bottoms of water. There are no limitations to depth in water except those to which plants grow and are agitated by waves or currents. Chances of preservation are good for the markings made on bottoms of water bodies, and as similar markings must have been made in the past it is probable that structures of this origin are present in the geologic column.

Counterparts of these markings may be made on the undersides of overlying strata.

TRACKS AND TRAILS OF ANIMALS

Animals have made tracks since their first appearance and such marks are known in the geologic column since the Cambrian. Trails made by worms, or wormlike and other animals, are extremely abundant in parts of the geologic column.

Closely akin to trails are burrows made by worms and wormlike and other animals. Burrows commonly have small elevations surrounding openings at the surface, and they may have vertical, horizontal, or any intermediate position. Some burrows are U-shaped and have two openings at the surface. Occasionally only fillings of burrows are found. These may be simple cylinders, as in *Scolithus*, or may have shapes like that of *Daemonolix* from the Tertiary of Nebraska.

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ICE-CRYSTAL IMPRESSIONS

The freezing of wet mud under certain conditions leads to aggregation and freezing of the contained water to form needle-like ice crystals. Freezing of wet sands produces similar results, but crystals are larger. Crystals are better developed in completely saturated than in partially saturated sediments. The ice needles or crystals approximate 1 to 2 mm. in length but may be up to 4 cm. long. Maximum widths are 1 to 2 mm., and many are not wider than a fine thread. Orientations on a small surface may be in many directions. They may be isolated in irregularly intersecting individuals or arranged in bundles or in fan shape radiating from a point. Most are straight; a few are curved. Melting of crystals leaves fine grooves that have the appearance of minute mud cracking.

Most ice crystals form in muds, and as these break down shortly after exposure the impressions are not likely to be preserved. If muds containing ice crystals or ice-crystal impressions become covered with sand, counterparts may be made, and these may be preserved. Markings referred to ice-crystal impressions have been observed in the Dakota sandstone of South Dakota and the Eagle Ford shales of Texas (Udden).

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RAIN, DRIP, AND HAIL IMPRESSIONS

Rain drops produce circular to elliptical impressions on soft muds and sands. The impressions are margined by slightly elevated, minutely ragged rims and have minutely rough surfaces. Rims are slightly higher on the sides toward which the forming drops were directed. Depths of impressions depend upon the velocity of the rain drop and the characters of the muds and sands, and impressions are deeper and have higher rims in muds than in sands. Maximum depths are about 3 mm.; widths range from about 2 to about 12 mm. Many impressions produce surfaces covered with coalescing pits (Fig. 79).

Rain impressions have been stated to be present on many horizons in the geologic column, but impressions more or less similar to those made by rain are made in many ways, and it is doubtful if rain impressions are as abundant in the geologic column as geologic literature indicates.

Drip and spray and splash impressions are like those made by rain. It is not known that any have been preserved.

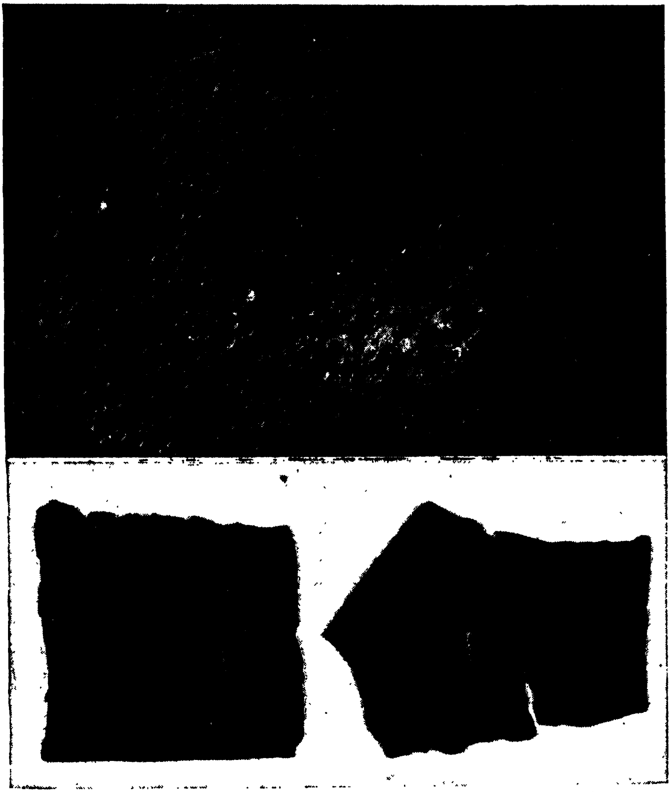

Fig. 79. Bubble impressions above; rain-drop impressions below. Made in the Sedimentation Laboratory of the University of Wisconsin.

Hail may make larger and deeper impressions than those made by rain, and some should be very deep and large, considering that hail as large as grapefruit has fallen and hail 2 cm. or more in diameter is common. Impressions made by hail should be common in the column, but beyond a possible occurrence in Triassic red shale of New Jersey none has been recorded.

With the exception of impressions made by hail, which may be made on the bottom of shallow water, the other impressions require exposed surfaces for formation, and hence occurrences indicate such a condition.

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PIT AND MOUND STRUCTURES

Pit and mound structures are made when muds settle rapidly from suspension. The settling muds acquire a certain rigidity at the surface, and small fissures form through which liquid from the interior finds egress. The fissures enlarge toward the top and close at the bottom. The concentration necessary for formation of the small fissures increases with coarseness of the muds. The rising liquids carry small particles in suspension and deposit some of these around the openings. This deposition forms small mounds, which are ultimately built to about 1 mm. high and from 3 to 10 mm. in diameter. Each has a small craterlike opening on the summit with a diameter of ½ to 1 mm. *Cross sections of laminated sediments, in which the pit and mound structures form, may strikingly resemble landscape marble,* and it is not unlikely that some of that structure developed in this way.

Pit and mound structures are thought to be possible of development in any depth of water, but, as an abundance of very fine-grained suspended sediments seems necessary for formation, it seems probable that the most favorable environment for formation is that of deltas and shallow sea bottoms repeatedly stirred by waves with consequent placing of vast volumes of finely divided sediments in suspension.

Pit and mound structures are known only from laboratory experiments, and they seem to form in any fine-grained sediment that is deposited from a thick suspension. The sediments may be powdered limestone. It is probable that they are present in the geologic column, although none has been identified.

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BUBBLE IMPRESSIONS

Bubbles that form in water may become stranded or be brought to the bottom through deposition of mud on the surfaces and may remain there as mud settles around them. Gases collect in sediments owing either to trapped air or to products of fermentation. These gases may rise to the surface, where they either remain for a while as sediments settle around them or escape upward through the water. Dislodgment of a bubble leaves a small depression of the dimensions of the part of the bubble buried. This has a smooth surface, is not margined by a raised rim, and may pass downward into a tube. These tubes bear considerable resemblance to the tubes made by burrowing worms or wormlike animals. Bubbles floating on the surface of very shallow water so that they are in contact with the bottom may make a very shallow trail greatly resembling those made by worms or wormlike animals (Fig. 79).

Pits due to stranding of bubbles are possible only in very shallow waters, but pits and tubes due to gases escaping from sediments may form on bottoms of any depth.

No bubble impressions are known in the geologic column, but it is not unlikely that they are present and have not been recognized or have been identified as rain-drop impressions, from which it is unlikely that fossil forms would show differences.

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Twenhofel, W. H., Impressions made by bubbles, etc., *Bull. Geol. Soc. Am.*, 32 (1921), 359–371.

CHAPTER XV

TEXTURES AND COLORS OF SEDIMENTS

TEXTURES OF SEDIMENTS

Textures of sedimentary rocks may be grouped into the five classes of fragmental, crystalline, spherulitic, oölitic, and pisolitic. Each of the first two may be subdivided, and there are all gradations between each and the others.

FRAGMENTAL TEXTURES

Fragmental textures are very fine grained (clay and silt textures), sandy, conglomeratic, bouldery, and brecciated. There are all gradations from any one to each of the others. These textures are generally considered to be confined to clastic sediments, which include claystones, siltstones, and the sequence of sandstones to conglomerates, but they are present in many limestones.

CRYSTALLINE TEXTURES

Crystalline textures are found in those rocks whose particles acquire shape by crystallization. Many limestones, nearly all evaporites, a few sandstones, and some mudstones have crystalline textures. In addition, many rocks of fragmental texture contain nonfragmental crystalline particles. Sandstones of crystalline texture are those whose particles have acquired crystalline shapes through enlargement.

Most sedimentary rocks placed in favorable conditions acquire crystalline texture. For many this requires heat greater than that normal to the surface, but sediments composed of carbonates, evaporites, and quartz more or less readily acquire crystalline textures under surface conditions. Time is naturally a factor, but favorable conditions must exist if time is to be effective. Dimensions of particles are probably greater for a longer than for a shorter time, and dolomitic limestones in a well at Spur, Texas, showed larger crystals with depth, suggesting, but not proving, that crystal dimension in these dolomitic limestones is a function of age and extent of burial. Limestones marginal to ancient coral reefs very commonly have macroscopic crystalline textures. Such are some of the crystalline limestones of the Chicotte formation (Silurian) on Anticosti Island and the so-called Hoburgen marble (also Silurian) on the Island of Gotland in the Baltic Sea. On the other hand, the limestones marginal to the Silurian coral reefs of eastern Wisconsin are fine-grained crystalline with crystals no larger than those in the limestones of the reefs.

Crystalline textures are microscopic in cherts and flints, fine-grained crystalline in most evaporites and many limestones, and coarse-grained in many of the same rocks. There are all gradations.

SPHERULITIC TEXTURES

Some sediments contain particles of spheroidal and ellipsoidal shapes and amorphous or radiate internal structure. These are termed spherulites, and the sediments in which they occur in considerable quantity have spherulitic texture. The particles resemble oölites and no doubt have been mistaken for them. Spherulites with radiate structure are evidently a form of crystal aggregate, whereas those with amorphous internal structure are either excremental particles or colloidal aggregates as glauconite, greenalite, or chamoisite. Spherulites with radiate structure seem to be largely calcite but many are pyrite or marcasite. Excremental particles have a wide range of composition, but there is commonly a large content of calcite.

OÖLITES AND PISOLITES AND OÖLITIC AND PISOLITIC TEXTURES

Introduction

Oölites are small spheroidal and ellipsoidal particles with concentrically laminated internal structure. Maximum diameters may be placed at 2 mm. Particles of the same shapes and same internal structure larger than 2 mm. in diameter are termed pisolites.[1] Rocks largely composed of oölites and pisolites have oölitic and pisolitic textures.

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[1] R. L. DeFord and O. A. Waldschmidt [*Bull. Am. Assoc. Pet. Geol.*, 30 (1946), 1587–1588] recommended that the terms oölite and pisolite be names of rocks and oölith and pisolith be names of the composing particles, the former less than 1 mm. in diameter, the latter 1 to 10 mm. in diameter. P. E. Cloud [*Univ. Texas, Pub.* 4301 (1945), 136] suggests oöid as the name

Many laminae of oölites and pisolites have radiate structure. Such is the case in some oölites that form in Great Salt Lake (Figs. 80 and 81) and in artificially formed oölites to be later described. Laminae may be indicated by different colors or different states of crystallization. Nuclei seem to be present in most oölites and pisolites. These consist of pieces of shells, fragments of many minerals, euhedral crystals of quartz (Henbest), pipe scale in some artificial oölites, and it seems probable that anything will serve. Gas bubbles have been suggested and oölites in some Mexican lakes are said to have had fish eggs as nuclei.

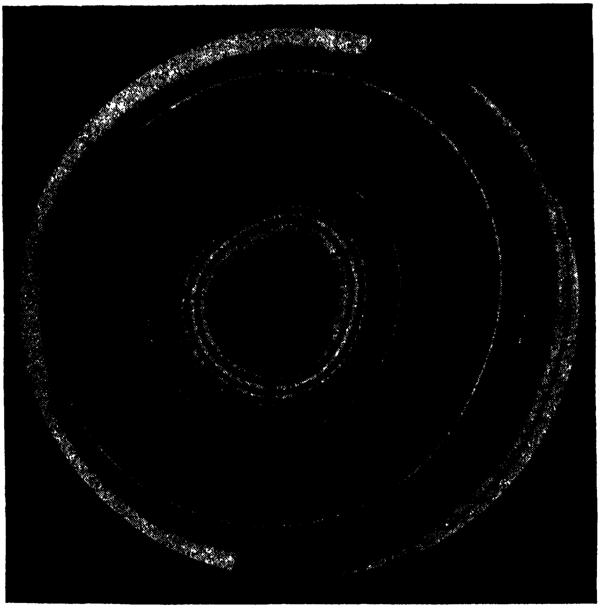

FIG. 80. Photomicrograph of an oölite "cave pearl," from the Cave of the Mounds, Dane County, Wisconsin. About forty times natural size.

A few occurrences of polyhedral pisolites have been described. These have the characteristic concentric internal structure, which on the outside terminates against the polyhedral faces.

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of the smaller particles. The author dissents from both proposals. The rocks composed of these particles are limestones, cherts, etc., and these rocks should be described as having oölitic or pisolitic textures.

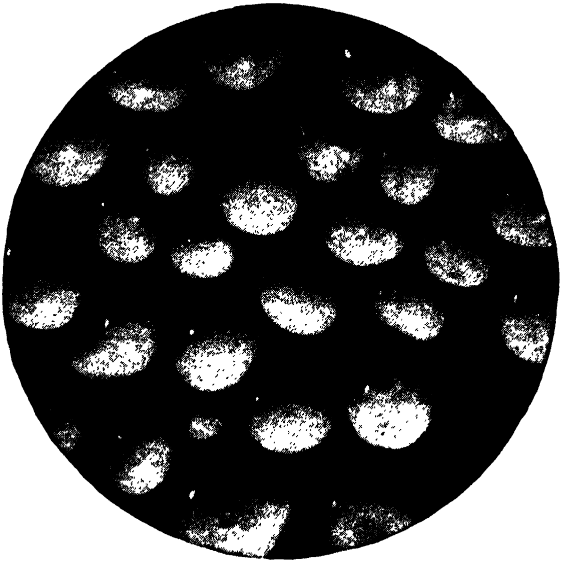

FIG. 81*A*. Photomicrograph of oölites with smooth surfaces. Some have mottled surfaces. From southeast shore of Great Salt Lake. Enlarged 15 diameters. (*Photograph by A. J. Eardley.*)

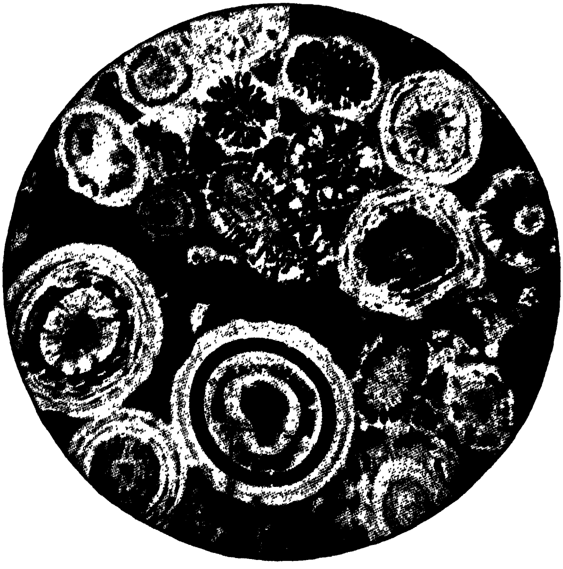

FIG. 81*B*. Photomicrograph of thin section of cemented oölites from Great Salt Lake. The more or less complete radiate and concentric structures are shown. The spherical particles with radiate structure are interpreted as fecal pellets. Enlarged about 35 diameters. (*Photograph by A. J. Eardley.*)

It is assumed that the polyhedral shapes are due to interferences of the pisolites with each other as they formed.

Oölites are present in the geologic column from Pre-Cambrian to the present. Some formations contain them in great abundance, as parts of the Mississippian strata of the Mississippi Valley and the Jurassic of England, the latter once being known as the *Oölite Formation.* In some formations, as the Oneota limestone of the upper Mississippi Valley, the oölites are confined to certain beds that are separated by other strata without oölites.

Composition of Oölites and Pisolites

Oölites and pisolites are known to be composed of calcite, aragonite, dolomite, hematite, limonite, pyrite, bauxite, chert and flint, barite, phosphate, and siderite. Some of these are known to be original and others are replacements.

Oölites and Pisolites of Carbonates. The most common carbonate in oölites and pisolites is calcite, and many beds and even formations are composed of calcite oölites, and some have many calcite pisolites. Calcitic oölites are now forming in some salt lakes, some parts of tropical seas, and some springs. They are known to form in both fresh and salt waters, but most seem to have been formed in salt water. Oölites composed of dolomite are present in many dolomitic limestones. These are thought to be replacements of oölites originally composed of calcite. Oölites of siderite are rare, and it is not known whether the few described are original or replacements.

Oölites and Pisolites of Silica and Silicates. Many nodules and beds of chert and flint contain oölites composed of silica. These are considered replacements of calcite oölites. Oölites of two dimensions, one with diameter of about 0.6 mm. and the other having diameters of 0.1 to 0.2 mm. in red and green shales in "Red Beds" near Lander, Wyoming, are considered original by Tarr. Oölites of light-green silicate with maximum diameter of 0.07 mm. are said to be present in Devonian limestones of Bath County, Kentucky, and oölitic texture has also been reported in fire clay at the base of the Pottsville in eastern Kentucky and in the St. Peter sandstone of Illinois. It is probable that these structures are original. The chamoisite in the Minette ores of France and in Wabana ores of Newfoundland has oölitic structure. It is not definitely known whether this is primary or a replacement.

Oölites and Pisolites of Iron Oxide, Hydroxide, and Sulphide. Oölites of iron oxide and hydroxide are extremely common, and the iron oxide deposits of the Clinton type are to a considerable extent composed of these. These are generally considered primary. Oölites and pisolites of iron oxide are forming in the so-called *shot soils* of the southern United States and elsewhere, and iron oxide pisolites are known to form on the bottoms of some lakes. The iron formations of the Lake Superior Pre-Cambrian contain oölites composed of concentric laminae of chert and iron oxide, or of red jaspery and black graphitic material. Some of the oölites have cleavage planes suggesting replacement of carbonate.

Oölites composed of pyrite are present in the Wabana ores of Newfoundland, the Devonian of Westphalia, and the upper Lias of northwest Germany. The pyrite oölites of Newfoundland are said to be primary.

Oölites and Pisolites of Phosphate. Oölites and pisolites of phosphate are characteristic of the Rocky Mountain phosphorites. Some contain spherulites as nuclei. Dimensions range from very small to 6 mm. in diameter. Opinion is divided as to whether the oölites and pisolites are primary or replacements.

Oölites and Pisolites of Bauxite. Most bauxite deposits contain oölites, pisolites, and larger concentric structures. The range in diameter is from microscopic to about 25 mm. These structures are primary and developed in the process of weathering.

Oölites and Pisolites of Barite. Oölites and pisolites of barite of primary origin formed in oil wells in the Batson and Saratoga pools of Texas. The fluid in the wells consisted of oil and water with a little sulphuric acid and had a temperature of about 52°C. Nuclei consisted of fragments of pipe scale, and dimensions ranged from 1.25 to 5 mm.

Origin of Oölites and Pisolites

Oölites and pisolites are probably present in the rocks of the column under two conditions, under one of which they occur in the environment of formation and in the other have been transported from the environment of formation into another. They may be transported as sands and pebbles and become ripple-marked and cross-laminated and be thrown upon shores where they may be picked up by winds and transported as dune sands, as is now being done on the shores of Great Salt Lake, the Red Sea, and the coast of Florida. Oölites and pisolites are known in strata that show little or no evidence of wave and current action. It is thought that these occurrences may represent the environments in which the oölites formed.

Pisolites have been seen forming in the Carlsbad springs of Bohemia. Nuclei consist of small particles of quartz and feldspar which receive coatings of calcium carbonate as they are kept in suspension and rotated by the bubbling waters of the springs. Oölites and pisolites have been seen to form in small pools on the floor of Carlsbad Caverns (Hess), New Mexico, and Holton Cave (Keller) near Columbia, Missouri. Drip from the roof agitates the water in the pools, and the agitation rotates nuclei and growing oölites and permits escape of carbon dioxide. Escape of carbon dioxide leads to precipitation of calcium carbonate, of which some is deposited upon nuclei and oölites. The pisolites range in diameter from 5 to 25 or 30 mm. Structure is concentric, and some laminae have radiate structure. Oölites have been formed in mines at Lead, South Dakota, in a mine in Idaho, in Australian caves and mines (Baker and Prostick) and in some Mexican mines under conditions somewhat similar to those existing in Carlsbad Caverns, except that rotation of the particles is stated not to have taken place (Davidson and McKinstrey). Other occurrences have been described by Mackin and Combs and by Pond.

Oölites of concentric structure with radiate structure in the laminae formed in the hot-water coil of the writer's furnace. These have nuclei of pipe scale, and there must have been considerable agitation of the water during their formation. The waters were also hot. At Seymour, Texas, water spray from an ammonia-cooling tower of an ice plant formed oölites without radiate structure around the base of the tower (Kemp). Nuclei consisted largely of sand, but one or two were coal. Diameters were mostly less than 2 mm. The oölites were composed of calcium carbonate, supposedly precipitated through escape of carbon dioxide caused by warming and agitation.

It does not seem probable that organisms were concerned in the formations of the oölites noted above. Except for the occurrences in Mexican and South Dakota mines, there was rotation of nuclei and the forming oölites, this permitting calcium carbonate to be deposited on all sides.

The formation of oölites and pisolites in and about Great Salt Lake and the Red Sea was ascribed by Rothpletz to algae. Oölites in the Jurassic of England and in salt lakes of the Kalihari Desert have also been referred to algae, and it has been stated that algal filaments are present in some. Linck's studies led to the view that all oölites and pisolites are of inorganic origin, and that oölites may form from precipitation of calcium carbonate around centers, either as aragonite or calcite, depending upon conditions and causes of precipitation. Studies of gallstones by Schade led to the view that oölites and spherulites are formed from deposition of colloidal materials, and that the precipitates assume spherical form that has radiate structure if the composing precipitates are pure, but concentric structures if impurities become enmeshed in the growing structures. Oölites in the Tertiary Green River formation have concentric but no radiate structure (Bradley). The concentric laminae are very high in ferric oxide, which is very low or wanting in the cementing matrix. Bradley postulated that the oölites formed in a gel of ferric hydroxide containing much finely divided calcium carbonate which may also have been in colloidal form. Coagulation of the ferric hydroxide by negative ions of carbonate or chloride led to coalescence of the gel into spheres that enmeshed the calcium carbonate suspended in the mixture. Dimensions were determined by the quantity of ferric hydroxide, and growth stopped when the supply was exhausted, leaving the remaining carbonate to form the matrix. Organisms are not considered to have been concerned in the formation.

According to Mathews, the Great Salt Lake oölites begin their history in shallow water, where one or two concentric laminations may form on nuclei. After being washed upon mud flats, they grow as blown inland by wind. The laminae correspond to seasons and result from precipitation of calcium carbonate through evaporation of water that was brought to the surface of the oölites by capillary action. Growth takes place mainly during early summer months when the temperature is highest, and little if any growth takes place during the winter. Soot is considered to have been deposited on the surfaces after the rainy season. Only 4 of 574 oölites examined were found without solid nuclei.

Eardley, the latest student of the oölites of Great Salt Lake, disagrees with the views of Mathews and was not able to find the soot rings described by the latter. He finds that some nuclei are fecal pellets, but most are quartz particles, and others are orthoclase, calcite, and a considerable variety of heavy minerals. Some oölites have radial structure in the laminae, and these are composed of calcite. Nonradiate oölites have opaque laminae, which appear dark in transmitted light, and it is suggested that these are the soot bands of Mathews. These bands seem to be composed of aragonite and are considered to be a primary precipitate from the alkaline waters of the lake. According to Eardley the oölites are found only along shore lines, and are most plentiful on shores exposed to large waves, and the more vigorous the wave action, the larger and more typically developed are the oölites. Eardley postulates that the carbonates constantly precipitated in Great Salt Lake over the areas of oölite formation attach themselves to nuclei, which in agitated waters are rolled over and over so that deposition is made on every side. It is suggested that the time of greatest formation is in the period from June to November.

It is considered probable that the structures designated oölites and pisolites form under various conditions. The "shot" in laterites show that formation does not require a water cover, but it seems certain that water must be present. It seems likely that the ferric and aluminum hydroxides that form oölites and pisolites in residual products were precipitated as colloids. Although it is possible that the calcium carbonate of some oölites was precipitated from materials in colloidal form, the actual formation where it has been observed shows that oölites of concentric structure with radiate structure in the laminae may form from precipitation of materials in solution. Proof seems wanting that oölite formation is due to algae, although it may well be the case that algae and other organic precipitants of finely divided calcium carbonate may be indirectly responsible. The inclusion of algal filaments may be incidental and not causal. Most calcite oölites formed in salt waters, but this is not a necessary factor for formation, as they also form in fresh waters. Formation of calcitic oölites in salt waters may perhaps be connected with the fact that these waters are more nearly saturated than fresh. Calcitic oölites do not seem to form in very cold water, which may be due to difficulty of precipitation of calcium carbonate because of the ease with which carbon dioxide is held in cold waters and hence may be due to difficulty of saturation. Most are now forming in warm waters. Calcitic oölites are not known to form in deep waters, and none are known in sediments of the bathyal and abyssal bottoms. Most seem to be confined to shallow waters with much wave action, but some additional factors are evidently required, as they do not form in all shallow waters of the regions of formation. Deposits rich in algal materials often have oölites and pisolites in association, but not all deposits rich in algal materials contain oölites and pisolites. All that seems to be known is that calcium carbonate precipitated under some conditions is deposited as films around nuclei. It seems probable that no generalization has general application to all oölites, and that oölites of the same or different composition form as consequences of different combinations of conditions.

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COLORS OF SEDIMENTS

INTRODUCTION

Colors of sediments are primary, or original, and secondary. Primary colors are those present at the time of burial of the sediments; secondary colors result from changes that take place subsequent to burial. Secondary colors may be produced while sediments are beneath the surface or after they enter the zone of weathering.

It is often difficult to determine whether a color is primary or secondary, and in some cases it seems impossible. Many sediments that are black at the time of burial because of the presence of the black sulphides of iron become gray as these sulphides change to marcasite or pyrite. Other sediments that are black from contained organic matter may have the organic matter destroyed by bacterial action and become gray to white as a consequence. Red or brown sediments may have the coloring ferric oxide or hydroxide reduced by organic matter and become gray. Sands that are white at the time of burial may become red through oxidation of contained iron carbonate or sulphide some time after burial. Most evaporites are white or gray and they are apt to remain so. Such is also the case for many limestones. Primary colors are more important in reconstructing the environment of origin of sediments than secondary ones. Color should be determined from unweathered rock.

Colors of sediments have a wide range, and probably every color may be found. *Colors of wet sediments are different from those of the same sediments dry, and the colors in sunlight are different from those in shadow.* These differences complicate the problem. Few geologists are sufficiently well trained with respect to color to make accurate statements, and the brown of one may be the red of another. The result is that many statements of color mean little to readers. It is hence recommended that a color chart be used to which the colors of sedimentary rocks and sediments may be referred.[1]

For convenience of consideration, the colors of sediments have been placed in the four groups of white to light gray, dark gray to black, yellow to red, and green. There are no sharp divisions between these groups. The most common colors are thought to be gray, yellow, brown, red, and black. Many sediments that seem blue when wet become gray when dry. Statements that follow apply to dry sediments.

White to Light Gray

Most evaporites and many limestones, claystones, siltstones, and sandstones have colors in the range from white to light gray. If much muscovite is present, as may be the case with some claystones, siltstones, and sandstones, there is a silvery luster to the color. Some mudstones and sandstones are light to dark blue.

It does not follow that rocks with white to light-gray colors were derived from sediments of the same colors. Some of the parent sediments probably were light-colored, but others may have had shades of yellow, red, or even black. The colors of the rocks may throw considerable light on the nature of the environmental conditions that gave rise to the parent sediments, and effort should be directed to determination of whether the colors of the rocks are primary or secondary. This is not easy.

Dark Gray to Black

Colors in the range from dark gray to black are largely due to the presence of organic matter, the black sulphides of iron and oxides of manganese and intrinsically black minerals. One or more of these may be present. They are named in what is thought to be the order of importance as coloring agents.

Organic matter may be carbonaceous or bituminous. The former includes all carbonized organic matter, the latter includes asphalt, tar, and other dark hydrocarbon compounds. The degree of darkness of color depends upon the quantity of organic matter present. Sediments with color due to organic matter are deposited in most environments, but most dark sediments of which the coloring matter is carbonaceous are deposited under continental conditions, whereas most of those of which the coloring matter is bituminous are believed to have been deposited in connection with the sea.

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[1] Goldman, M. I., and H. E. Merwin, National Research Council, 1928. Goddard, Trask, DeFord, Rove, and Singewald. National Research Council, 1948.

Sediments deposited under conditions of a limited quantity of oxygen range in color from blue or gray to black. If organic matter is not present the color may be assumed to be due to black sulphides of iron. After burial the darkness of color due to these substances is lost, as the compounds are unstable and change to marcasite or pyrite. Rocks containing these minerals again change color when they enter the zone of weathering, as oxidation takes place, and the colors change to the range from yellow to red. Sediments of which the colors are due to the black sulphides of iron are deposited in such inland bodies of salt water as the Black Sea and Lake Baikal, deep holes in the bottoms of shallow seas, and bays and sounds of arms of the sea with little or no tide as in the east Baltic Sea. The development of the sulphides is favored by salt water and poor circulation.

Some other black minerals are deposited from solution, as the black oxides of manganese, which form black coatings on rock particles and shells over many parts of the sea bottom, some fresh-water bottoms, and the land also. Clays, silts, and sands may become black from this cause. Occurrences of dark colors due to manganese deposition are insignificant in comparison with other substances producing these colors.

Allogenic dark-colored minerals and rock fragments are perhaps third in order of importance in giving dark colors to sedimentary rocks. The colors are due to such minerals as hornblende, augite, biotite, magnetite, ilmenite, and fragments of such rocks as black slates, black flints, basalt, scoria, and obsidian. Sediments with dark colors due to the presence of these original minerals and rocks tend to accumulate under conditions wherein decomposition is limited and physical destruction of rocks pronounced. Short transportation may be generally assumed.

YELLOW TO RED

Sediments in the range of color from yellow to red are extremely common, but there is great difficulty in determining whether the colors are primary or secondary, as many sediments contain iron in the bivalent form, which readily oxidizes on reaching the zone of weathering and imparts secondary colors in the range from yellow to red. The primary colors may have been white, gray, black, green, or some other color. Most residual materials of tropical and tropical-temperate upland regions have color in the range of yellow to red, and, if sediments from these materials are deposited under conditions excluding organic matter, the primary color is retained, but if organic matter is present, the iron is reduced to the bivalent form, and the colors of oxidation are lost. Exposure to agents of weathering may reoxidize the iron with return to original color.

The red color is due mostly to hematite and the yellow to limonite. Intermediate shades originate from mixtures of the two minerals.

Some sediments are locally red because of color of such original minerals as garnet and red feldspar. Some sediments are also red from the presence of particles of red jasper, red slate, red quartzite, and red felsitic lava.

GREEN

Green is not a common color of sediments. Some green sediments contain a great abundance of serpentine, epidote, chlorite, and fragments of green slate. Others are green from the presence of authigenic glauconite and greenalite. Glauconite is probably the most important of the green minerals in sediments, and it gives color to many claystones, siltstones, sandstones, and some limestones. Some sediments contain so much glauconite as to merit the designation of greensand. Greenalite is a common mineral in some of the Pre-Cambrian rocks of the Lake Superior Region and gives green color to formations in which it is present in any quantity. Hornblende, uralite, bastite, and olivine are in some cases responsible for green sediments. Olivine is a common and in some places the most abundant constituent of Hawaiian beach sediments. Malachite in rare cases colors sediments.

SIGNIFICANCE OF COLOR

Color of sediments is influenced by all environmental conditions that are concerned with origin, transportation, and deposition. Among these are the character of the parent rocks; methods of rock destruction; climatic conditions; methods, distances and duration of transportation; conditions at places of deposition; and diagenesis subsequent to deposition. Color has greater significance in continental than marine sediments, as deposition of the latter is usually much slower and the quantity of organic matter buried with the sediments is usually adequate to produce changes of color of terrigeneous sediments after they reach the sea. The results are that sediments deposited under normal marine conditions are likely to have the same general colors in the range from gray or blue to green.

The significance of the parent rocks and the colors of weathered materials under different climatic conditions have been considered, and only summary statements are given here. These are as follows:

Sediments derived from parent rocks that have experienced little decomposition have the colors of the original minerals, and this condition exists in hot, rigorously arid regions; rigorously cold regions; regions of sharp relief and excessive rainfall; and on beaches backed by cliffs undergoing active erosion. If the rocks are ground to powder the colors tend to be light, as rock powders are usually light colored. If the cold conditions permit vegetation, organic matter enters continental sediments in quantity, and darkness of color obtains.

If the climatic conditions are those of the polar halves of the temperate zones and are not arid, parent rocks are usually decomposed, vegetation is generally present to reduce the iron, and there is usually sufficient water to remove compounds of ferrous iron. The residual materials are light colored poleward and yellow equatorward. These sediments are transported and deposited with organic matter, oxidation of iron salts is prevented, reduction of iron salts is favored, and the color of the sediments is in the range from light to black, depending on the quantity of organic matter buried with the sediments and not destroyed by decomposition before burial.

In tropical and tropical-temperate upland regions of moderate to more rainfall, rocks are chiefly destroyed by decomposition. Destruction is usually complete, with all soluble products removed, leaving a more or less red residue of aluminum hydroxide, hydrous aluminum silicate, hydrous and anhydrous iron oxide, quartz, and a few other resistant minerals. All residual matter without organic matter is brown to red. Transportation of these residuals may mix them with organic matter with loss of the red color. If deposition takes place without admixture of organic matter, the red is retained and gives mute evidence of the producing environmental conditions (see "Red Beds").

In the sea the factors most important so far as color is concerned are the minerals produced by the marine conditions, which remain to tell of the conditions, and the diagenesis that is almost certain to take place by reason of deposition of organic matter with the sediments. Under many conditions, bacterial decomposition destroys all organic matter, and originally black sediments may become white to gray. Under some conditions glauconite forms and imparts a green color (see topic Glauconite). In the deep abyss the sediments attaining the bottom are thoroughly decomposed, the iron oxidized, and the organic matter destroyed in the long passage to the bottom. The red clays result. These are rare in the column, which is probably due to the rarity of elevation of deep-sea bottom.

Under anaerobic conditions, organic matter is not destroyed and remains to give color to sediments. Sulphate-reducing bacteria may throng in such environments and form the black sulphides of iron. Anaerobic conditions exist in waters of poor circulation as in deep holes in shallow marine bottoms, bays and seas with weak or no tides, and many lakes.

Sediments on shores and near shores feel the influence of local conditions, and the sediments have colors determined by the native rocks, physiography, and climate.

SUMMARY

1. Sediments with colors like those of the parent rock suggest rigorously cold or rigorously arid climatic conditions, extremely steep slopes, or beaches and near-shore bottoms of rivers, lakes, and the sea, or regions of moderate to sharp slopes with such excessive rainfall that a plant cover cannot develop.

2. Black in lithified sediments usually indicates conditions preventing complete decay of organic matter, and the conditions of marshes, wet and cool to cold floodplains, lakes in which there is imperfect overturn, shallow tideless or nearly tideless marine waters, deep holes in shallow marine bottoms, or lush accumulation of organic matter.

3. Gray sediments are indicative of no particular environment,

as they may be deposited in lakes, playas, floodplains and deltas, and in shallow and moderately deep seas. They may also be deposited in cool to warm climates.

4. Colors that change from bed to bed are suggestive of the deposits of alluvial fans, floodplains, and deltas, but this condition may also be found in marine sediments.

5. Green as a primary color denotes glauconite deposits of the sea or original minerals of parent rocks and, hence, indicates conditions not favoring decay. Glauconite is probably indicative of shallow water and slow to negative deposition. Green as a secondary color may be found in some pyroclastics.

6. Primary red colors indicate, first a warm climate, an upland region, and sufficient rainfall to support a good but not an abundant growth of vegetation at the places of production and, second, deposition under climatic conditions of hot and dry seasons alternating with rainy seasons, or general dryness. Deposition may take place on river floodplains and deltas of dry regions, or in shallow lakes and the sea, with high salinity preventing growth of organic matter. The red sediments may also have been derived from a terrane of red sediments or one composed of red minerals. The red clays of the abyss are not included in the preceding statements.

The simple statements made in this summary are intended to serve merely as signposts. In actual practice a geologist meets with many sediments of complex history. Some yellow sands on the La Crosse River floodplain of western Wisconsin were laid down by winds. The winds derived the sands from glacial outwash on the La Crosse River floodplain. The melt waters acquired the sands from glaciers, and the glaciers acquired them from the Cambrian sandstones that were laid down in the sea and were derived from some Pre-Cambrian terrane. What is the environmental significance of the yellow color? Some rivers originate in high mountains with arctic cold, flow through tropical forests with lush vegetation, pass through regions with arid conditions, and finally yield sediments to deltas and the sea. What is the significance of the colors of these sediments? It is difficult to state.

With colors, as with most sedimentary features, simple rules are impossible; suggestion should be the rule, and each color constitutes a problem. The color of a sedimentary rock may have significance for solution of the problem of the environment of deposition of the composing sediments, but it also may have no significance in that respect.

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